Institute of Polar Studies

Report No. 55

A Review of Quaternary Investigations in Greenland

by

Anker Weidick
Geological Survey of Greenland
in cooperation with
Institute of Polar Studies

October, 1975

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INSTITUTE OF POLAR STUDIES

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A REVIEW OF QUATERNARY INVESTIGATIONS IN GREENLAND

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Østervoldgade 10, DK-1350
Copenhagen K, Denmark

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Requests for this publication should be addressed to:

Geological Survey of Greenland
Østervoldgade 10, DK-1350
Copenhagen K, Denmark

EDITOR'S NOTE

This report was originally prepared by the author as a publication of the Geological Survey of Greenland (GGU). It represents many years of work on the Quaternary of Greenland, mainly by the author and his colleagues. In order to expedite publication, arrangements were made between the Geological Survey of Greenland and the Institute of Polar Studies to publish the manuscript in the Institute's Report series.

GGU graciously provided copies of their "Quaternary Map of Greenland" to accompany the Report. All requests for the Report should be directed to the Geological Survey of Greenland, Copenhagen.

Although the manuscript style and format did not necessarily conform with the Institute's Report series, the author's copy was published more or less as he wrote it, in order to preserve the style and geographic names usage of the Geological Survey of Greenland.

J. F. Splettstoesser
Lincoln, Nebraska

ABSTRACT

The Quaternary deposits of Greenland consist of two generally separately treated units: the present ice covers (essentially the Inland Ice); and the sediment deposits of the present ice-free coastal areas. In order to link together the present information of the two areas, the results of the recent investigations of the Inland Ice are treated in order to furnish a background for the Quaternary history of the coastland.

On the Quaternary history of the coastland, emphasis is placed on the glacial history, especially deglaciation during the Holocene. In connection with the Holocene history and the landscape features thus developed, a short account is given of the investigations of the present periglacial and other Quaternary phenomena (warm springs, sulfur mounds, meteorites).

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J. Splettstoesser reviewed and edited the manuscript. P.J. Anderson prepared it for publication. J. Cothran typed the final publication manuscript.

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* Dr. Ten Brink's present address is Department of Geology, Grand Valley State College, Allendale, Michigan 49401.

** Dr. Langway's present address is Chairman, Department of Geological Sciences, State University of New York at Buffalo, 4240 Ridge Lea Road, Amherst, New York 14226.

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INTRODUCTION

Greenland is situated near the center of the area covered by the great Pleistocene glaciations of the northern hemisphere, and the existing Inland Ice of Greenland is the largest and possibly the only relict ice of the ice ages here in the northern hemisphere. These two facts alone make an understanding of the glacial history of this island particularly desirable. However, these same facts make a reconstruction of Greenland glacial history highly speculative because most evidence of former glaciations either has been removed or is concealed by the present ice.

In order to stress the principle of uniformitarianism as a basis for this treatise on the Greenland Quaternary, the first section deals with the present topography and glaciology of Greenland and its surroundings. The next section treats the development of glaciation and is followed by a discussion of glacial and interglacial ages. Finally, the Holocene record of marine and glacial deposits is reviewed and followed by a brief discussion of vegetation, archeology, periglacial phenomena, warm springs and meteorites.

In the regional description of occurrences the district and province division of Greenland in general follows that given in the General Index of Meddelelser om Grønland (vols. 1-150, 1969, ed. A. Andersen), here shown in Fig. 40. An exception is that the administrative units of Narssaq, Julianehåb and Nanortalik in southernmost West Greenland is here labeled Julianehåb district due to a more natural, but older division. East Greenland is used as a collective term for North-East and South-East Greenland.

MAIN TOPOGRAPHIC ELEMENTS OF GREENLAND

A natural division of the gross topography of Greenland falls into the following units: the submarine areas, the ice-free coastal areas, and the subglacial topography of the present Inland Ice. A fourth unit: the present ice cover, is treated in the section on present glacial conditions, p. 15-47.

The submarine areas

A general outline of Arctic bathymetry is given by Stearns (1965) and much of the current literature on the subject has been summarized by Gakkel and Dibner (1967).

With reference to the bathymetry of Greenland waters, most parts off West and East Greenland are covered by Charts of the Royal Danish Hydrographic Office. This source can be supplemented for East Greenland by the results of the work by the L. Boyd expeditions (1935, 1948), Johnson and Heezen (1967), Laktionov (1959), Litvin (1968), and Svirenko and Soldatov (1964). In North-West Greenland the areas between the Arctic Ocean and Baffin Bay have been treated by Pelletier (1964). North Greenland waters are covered by a chart of 1:2,000,000 scale by the Canadian Hydrographic Office.

The deep sea

The bathymetric charts show Greenland to be surrounded by oceanic basins 2,500 to more than 4,000 m deep. The basins are mutually separated by ridges or rises, the most important of which connects Greenland with Iceland and the Faeroe Islands. The maximum depths above this rise are 500 to 600 m, and thus it forms a threshold for exchange of Atlantic and Arctic waters. This threshold is believed to have had a great influence on climatic fluctuations during the Pleistocene in connection with eustatic changes of sea level (Ewing and Donn, 1961).

Along the Lomonosov Ridge in the Arctic Basin, Pleistocene volcanism is thought to have occurred (Gakkel, 1958), but the adjacent Nansen Ridge is now known to be the continuation of the volcanically active zone that parallels the East Greenland coast from Iceland to Jan Mayen.

The shelf

The shelf around Greenland is 50 to 300 km wide, its greatest expanses occurring off Jøkelbugten in North-East Greenland and

Kangerdlugssuaq fjord in South-East Greenland (Fig. 1). In the latter area the shelf grades into the rise between Greenland and Iceland. Off West Greenland the greatest shelf area occurs near Disko Bugt. In North Greenland only sparse information is available but the shelf area seems to have a width of 50 to 100 km.

Water depths over the shelf are mostly between 100 and 300 m off West Greenland, North-East Greenland and possibly North Greenland. In South-East Greenland, from Scoresby Sund to Kap Farvel, the shelf is somewhat deeper, and wide areas occur 300 to 400 m below sea level.

Greenland is connected to Ellesmere Island through a common shelf between Baffin Bay, Smith Sound (west of Thule) and the Arctic Ocean. The surface of the shelf area in this region is dissected by a chain of channels and basins, which may be explained as being drowned drainage systems of Tertiary age, primarily tectonic, but modified by Pleistocene glacial action (Pelletier, 1964).

Sinuous troughs, 600 to 1,000 m deep, cut through the Greenland shelf at intervals and extend from fjord systems or depressions in the coastland to the continental slopes. The longest and deepest depression seems to cut across the shelf in Jøkelbugten, but the exact form and length cannot be given from the current information. The longest and deepest trough in West Greenland also seems to occur toward the north, in Melville Bugt, where a depression parallels the coast for a great distance.

For West and East Greenland there are two characteristic southward trends of the shelf area: a general decrease of width; and an increase in the presence of thresholds which act as barriers between the ends of troughs and the continental slope (Fig. 2). The troughs off West Greenland, as in Smith Sound, are thought to be drowned drainage systems of Tertiary age (Pelletier, 1964), and depressions in the shelf off East Greenland may be similarly explained. The increasing occurrence of thresholds toward the south cannot be explained simply by a lack of soundings in northern waters and may be due to differences in pre-glacial, and especially glacial, erosion and deposition.

Near the coast, the Greenland shelf is often furrowed by marginal channels that trend parallel to the coast. These are presumed to mark the limit between the continental landmass and relatively younger sediments off the coast, as in Norway (Holtedahl, 1970).

Surficial deposits of the shelf

Maps of surficial deposits off West Greenland by Rvachev (1963) cover the area between 59° and 68° N. Horsted (1969) also gives

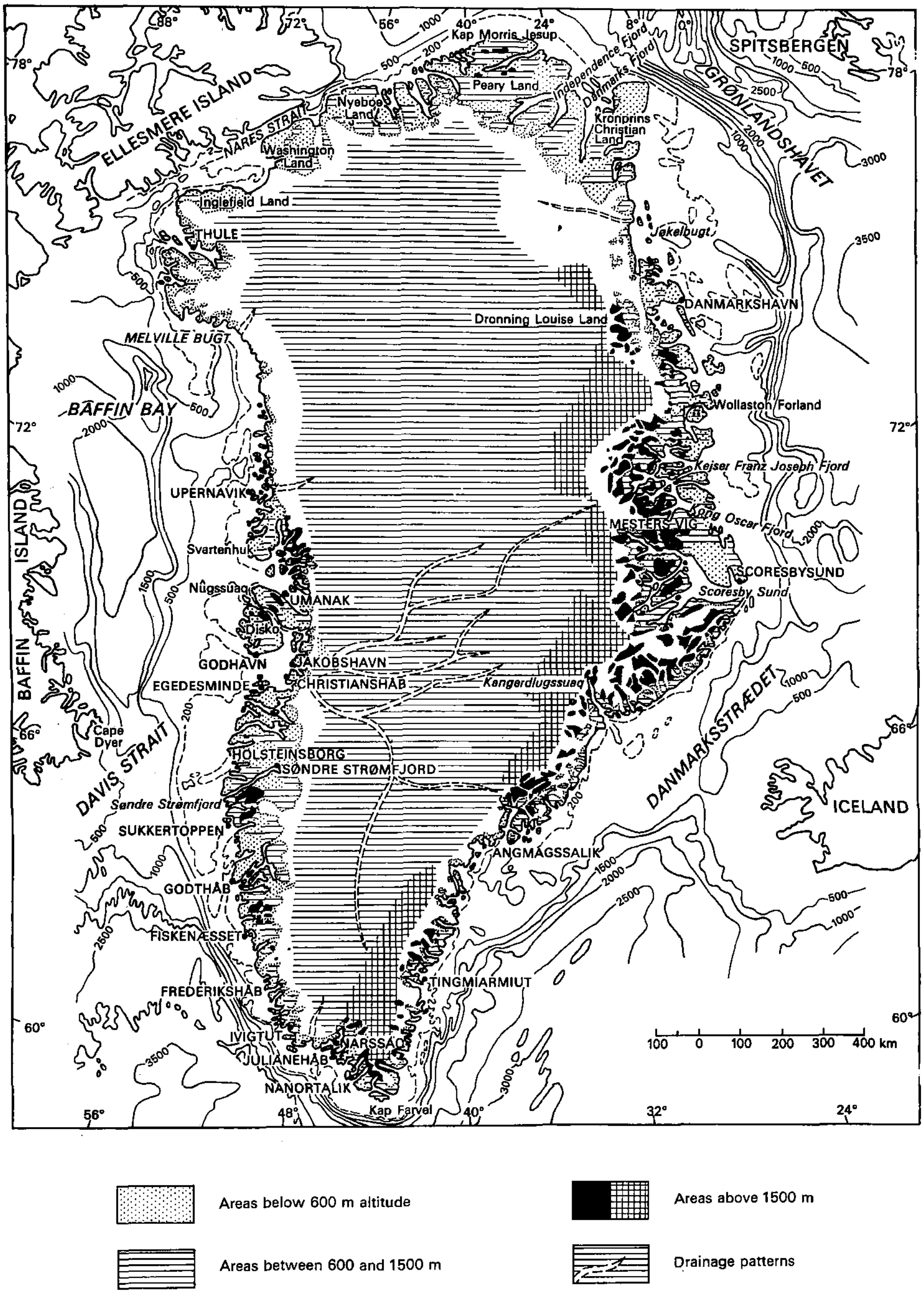


Fig. 1. Topographic map of Greenland and surrounding areas. Bathymetric curves after the sources mentioned in the text. Topography on the coastal stretch after the maps of the Geodetic Institute, Copenhagen, and World Aeronautical Charts. Topography of the inland areas based in measurements of Expéditions Polaires Françaises, U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) and the British North Greenland Expedition. Topography here shown corrected for the present ice load when Inland Ice is removed. Altitudes above 1500 m are shown black where ice free and cross-hatched when covered by the present Inland Ice.

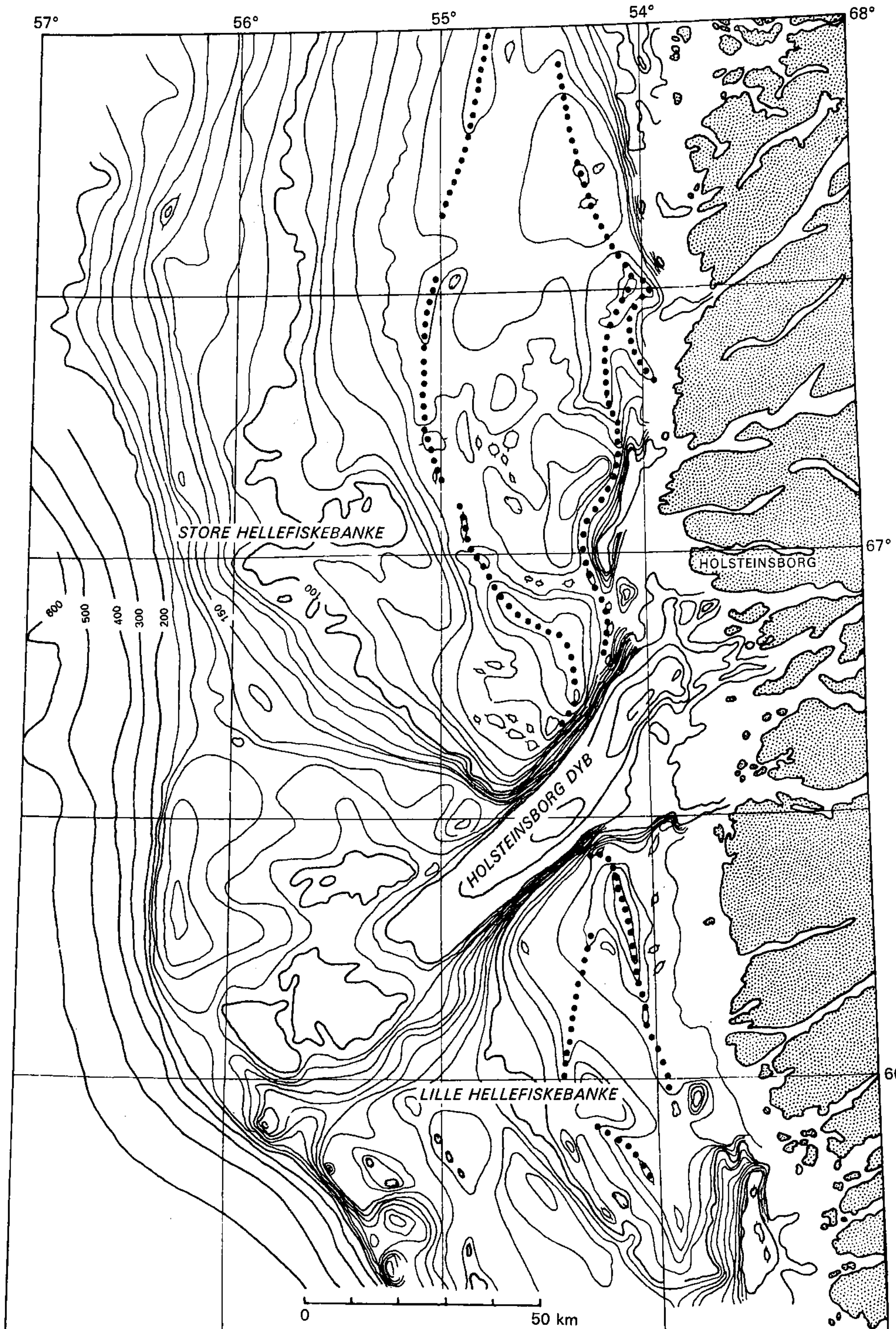


Fig. 2. Store Hellefiskebanke and Lille Hellefiskebanke separated by the Holsteinsborg Dyb (dyb = trough) off central West Greenland. Contour interval 100 m between thick lines and 10 m between thin lines. Dotted lines are submarine ridges, presumably end moraines. The Holsteinsborg Dyb is seen separated from the continental slope by three thresholds at approximately 200 m depth. Compiled from Charts 1401 and 1500 of the Royal Danish Hydrographic Office.

information about parts of the West Greenland shelf, expecially off Disko Bugt. The east coast of Greenland is covered from 59° to approximately 70° N by the International Quaternary Association Map of Europe 1: 2,500,000, sheets 1 and 2 (Woldstedt, 1967). North of these areas only very sparse information is available.

Large areas of the known shelf surface are covered with glacial marine sediments and boulder clay. Coarser sediments seem to be especially concentrated on the banks; i.e., the highest parts of the shelf. Off West Greenland, Rvachev shows the trough bottoms as covered essentially by clay.

The ice-free coastal areas

Topography

The topography of West Greenland from 59° to 78° N and East Greenland from 59° to 76° N is shown on Danish Geodetic Institute map sheets of 1: 250,000 scale. The ice-free areas in North Greenland are shown on International Civil Aviation Organization (ICAO) maps, scale 1: 1,000,000.

The very generalized topographic map (Fig. 1) delineates the areas between 0-600 m, 600-1,500 m, and more than 1,500 m above sea level. These limits are arbitrary but are believed to distinguish between highland nuclei of initial glaciations (above 1,500 m), uplands influencing drainage of the ice cover during the ice ages (600-1,500 m) and lowlands of the most important Wisconsin and Holocene deposition (below 600 m).

Limited coastal areas underlain by basalt and gneiss often exhibit typical alpine topography, but plateau-like areas are dominant, especially in the northern areas of sedimentary rocks. In the alpine areas, concordant summit levels reveal their origin from older peneplains.

Development of landscape

Most investigations of this subject have been made in East Greenland, where, after the breaking up of the pre-Cretaceous topography by a series of N-S trending faults, peneplanation took place in late(?) Mesozoic (Backlund, 1930) or early Tertiary time (Ahlmann, 1941). This peneplain is represented by summits and plateaus from 1,000 m above sea level at the coast to 2,000-2,250 m around the inner parts of Scoresby Sund. Because the late Mesozoic-Tertiary basalts of Scoresby Sund are truncated by these surfaces, Ahlmann concluded that peneplanation continued into the Tertiary, and that principal uplift of the denuded terrain also occurred during Tertiary time.

Peneplains of the same elevations as those in East Greenland were reported from southern West Greenland by Wegmann (1938). Similar erosion surfaces at lower elevations were reported from West Greenland by Nordenskiöld (1914) Birket-Smith (1928) and Paterson (1951). In the Umanak district, summit levels developed on metamorphic rocks have been treated by Drever and Wyllie (1951) and there is an apparent accord with a surface developed on basalts of the outer district. If the surfaces in these areas do in fact represent one peneplain, then they must have been leveled and elevated in the middle or late part of the Tertiary. Wegmann supposed this to be the period when uplift of the peneplain represented by summit levels in southern West Greenland occurred, an idea later worked out in detail for the island of Sermersôq by Oen (1965).

In North Greenland, summit levels of 300-400 m (Davies et al., 1963) are thought to accord with levels of the Carey Øer (Bendix-Almgreen et al., 1967), and possibly with surfaces in Inglefield Land from about 200 m at the coast to above 1,000 m near the Inland Ice (Nichols, 1969). The age of this peneplain is believed to be Tertiary or Pliocene - early Pleistocene (Davies et al., 1963).

In general, there is an apparent agreement of a Tertiary age for many high Greenland peneplains. Stratigraphic and erosional evidence led Wager (1933) to estimate that major uplift of Greenland peneplains occurred near Miocene time. Closer examination may reveal more cycles of peneplanation similar to those determined by Oen (1965) on Sermersôq in the Julianehåb district. Also in southern West Greenland (Frederikshåb district), the uppermost peneplains cut dikes of presumed Tertiary age. However, recent radiometric age determinations indicate a Cretaceous age for these dikes (Larsen, 1966).

The investigations summarized above include most of the highest plateaus in Greenland. Secondary, lower peneplain remnants are also known from many areas, and may be explained as being either remnants of dissected peneplains or younger surfaces, formed during secondary erosion cycles. The lowermost surface, the strandflat, is thought to be Pleistocene in age and is discussed on p. 56.

The subglacial topography of the present Inland Ice

Topography

Since the first determinations of ice thickness during the Wegener expedition (Brockamp et al., 1933), extensive work has been done in mapping the bottom topography of the Inland Ice. Considerable parts of southern Greenland were mapped by Expéditions Polaires Françaises (Holtscherer and Bauer, 1954), and more northern areas have been investigated by the British North Greenland Expedition (Bull, 1957) and the United States Army (Clarke, 1966; Roethlisberger, 1959, 1961b).

Most ice-thickness data have been obtained by seismic investigations, but gravimetric techniques were also employed in some of the traverses. The newest technique is radar echo sounding, which was initiated by Evans and Waite (Rinker et al., 1966), and airborne radar sounding is rapidly becoming more widely used (Gudmandsen, in Evans et al., 1969). The published results of radar echo soundings seem to agree with previous investigations (Walker et al., 1968).

Discrepancies between results from the investigations mentioned above amount to differences in elevation of the substratum of 100 to 200 m, which may be due to differences inherent in the applied techniques or errors in exact determination of position. In any case, these errors do not influence the general form of subglacial topography given in Fig. 3. This topography has been corrected for removal of the present ice load in Fig. 1 by the addition of a third of the overlying ice thickness to the measured values. With due reservations about present isostatic equilibrium of the coastland, Fig. 1 suggests that the pre-glacial landscape was essentially an inland continuation of the high coastal plateaus. This prompts speculation about a possible Tertiary origin for the subglacial landscape similar to that of the coastal plateaus, an idea made somewhat more tenable by: (1) the existence of a subglacial drainage system of deep valleys which converge toward Disko Bugt (Figs. 1 and 2), and (2) the probability that the source of late Cretaceous and early Tertiary sediments in the coastal area must be found under the present Inland Ice (Koch, 1964).

Minor subglacial valleys may have various orientations. For example, the Qagssimiut depression of Wegmann (1938) leads to the south from the southernmost part of the Inland Ice and crosses the present main fjords in the Julianehåb district at right angles. The age of this feature was considered to be Mesozoic or Tertiary by Wegmann.

For the northern part of Greenland, no exact evidence of subglacial drainage patterns exists. Free of ice, the area would seem to be a hilly plateau between 500 and 100 m above sea level. The deep depressions of Melville Bugt and Jøkelbugten may be continuations of subglacial drainage patterns under the present Inland Ice; however, this cannot be stated from the measured profiles of the substratum.

On the basis of coastal elevations and the existence of two Inland Ice domes Koch (1928a) states that a tectonically conditioned depression crossed Greenland between Disko Bugt and Kangerdlugssuaq. As seen in Fig. 1, there is no break in the eastern N-S trending highlands of the dimensions assumed by Koch. However, the fact remains that some of the main drainage patterns seem to continue the direction of Koch's presumed faulting zone.

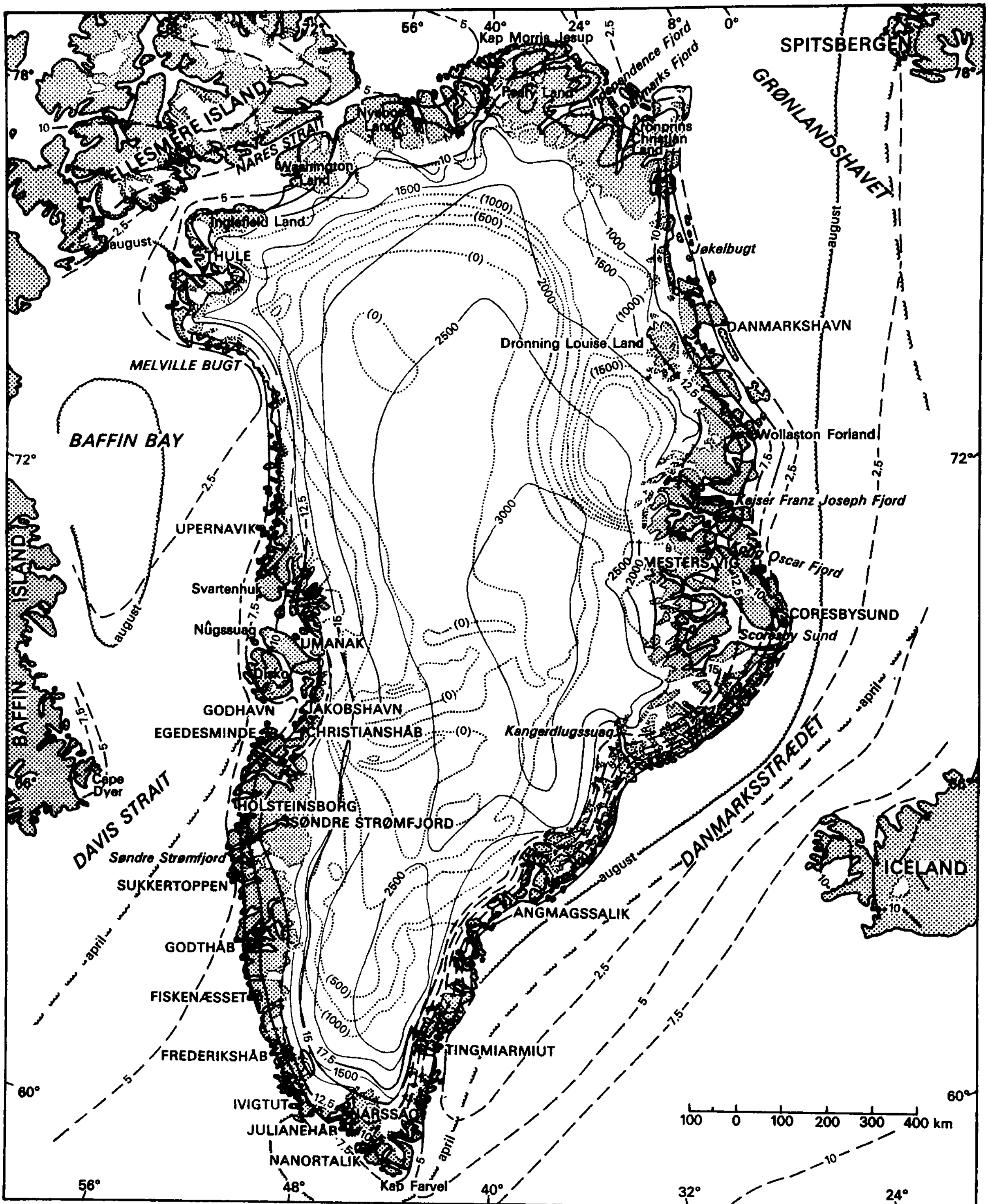


Fig. 3. Present glacial conditions of Greenland. Extent of drift ice in April and August gives the approximate maximum and minimum extents respectively. Generally the maximum extent occurs earlier on the east coast then on the west coast of Greenland. Over the coastland are shown isoglacihypses and over the margin of the Inland Ice firn line altitudes (thick lines). Surface contours of the Inland Ice (thin lines) after aeronautical maps and sub-surface contours (dotted lines) after the same sources as in Fig. 1.

Deposits under the present Inland Ice

Only at a single place, Camp Century at 77° 10' N, 61° 08' W, has drilling penetrated the Inland Ice (Fig. 4). The site is situated 1,885 m a.s.l., and the core has a length of 1,390 m and a diameter of 12 cm. The drilling penetrated 3.6 m of the Inland Ice substratum which was described as frozen till (Langway, 1967).

On the basis of seismic investigations, Holtzscherer (Holtzscherer and Bauer, 1954) supposed with some reservation that the substratum of the central part of the Inland Ice consisted of moraine, 200-300 m thick, which thinned out toward the ice margin. The observed velocity of propagation of seismic waves in this layer was 4,800-5,000 m/sec, and Holtzscherer compared this with similar velocities observed in fresh moraine in Alaska. The profile of the lenticular permafrost layer or frozen moraine is given in Haefeli (1959, fig. 3, p. 5). Temperature in the uppermost central part of this permafrost zone is supposed to be -10°C (Holtzscherer and Bauer, 1954).

Later investigations by Roethlisberger (1961a, b) on the propagation of seismic waves in different rock types of the Thule area showed that black shales and sandstones as well as outwash all have the same velocities as given above for frozen moraine. The temperature immediately under the active layer in the area is approximately -10°C.

In regard to older deposits, both the seismic investigations and the occurrence of erratic sandstone boulders in West Greenland (Steenstrup, 1883c; Barton, 1897; Rosenkrantz _et al._, 1942) suggest a possible subglacial continuation of Precambrian sedimentary rocks between outcrops in Julianehåb district and the Thule area. The probable continuation of Precambrian formations under the Inland Ice in southern West Greenland was mentioned by Noe-Nygaard, and is also implied by the occurrence of rapakivi in the Egedesminde district and boulders of rapakivi in the Umanak district (Noe-Nygaard, 1944; Pulvertaft, personal communication).

The West Greenland area of Cretaceous and Tertiary deposits also supposedly continues northward beneath the Inland Ice, because basalt boulders have been observed in the northern part of Upernavik district by Ryder (1889) and Carlson (1941). From the same district, Tarr (1897b) and Paterson (1951) reported erratics of slate and mudstone. A southern continuation of deposits of the same age under the West Greenland banks has also been postulated (Rvachev, 1963; Johnson _et al._, 1969).

In North-East Greenland, surveys of erratic boulders near the Inland Ice margin have been made by Parkinson and Whittard (1931), Backlund (1932), Teichert (1933), and Flint (1948). They all found the lithologies of erratics to correspond with those of outcrops near the ice margin, indicating a short distance of transport.

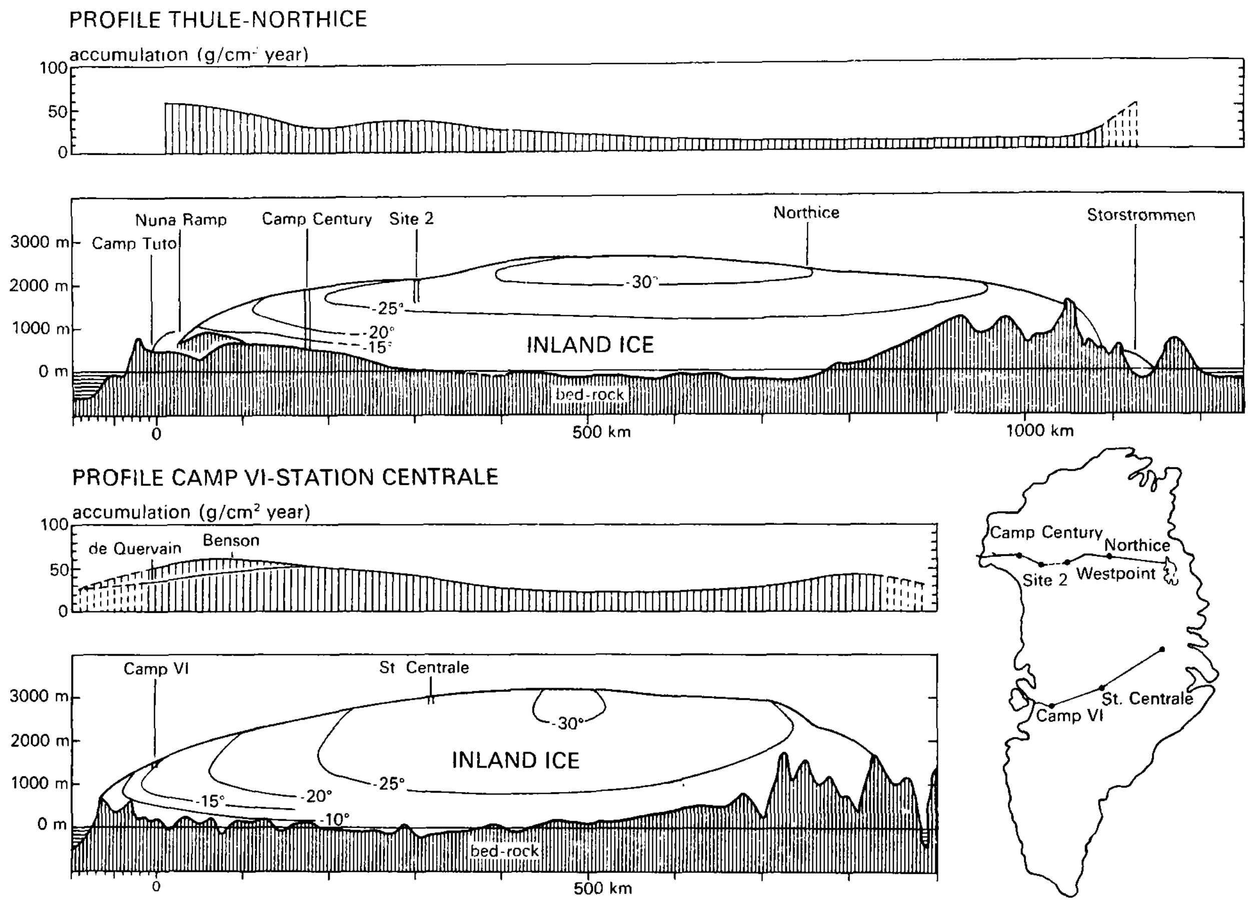


Fig. 4. Profiles of the Inland Ice across North and central Greenland. Positions of the sections are shown on the sketch map to the right. The thickness of the Inland Ice is shown on the profiles together with estimated temperature conditions on the Inland Ice and the depth of the core holes at Camp Century, Site 2, Camp VI and Station Centrale. Above each profile is shown accumulation of the same area. Data for northern section: accumulation, Benson (1962, fig. 29, p. 38); surface temperatures, ibid. (1962, fig. 37, p. 54); internal temperatures, Hansen and Landauer (1958, fig. 3, p. 316) and Weertman (1968, fig. 7, p. 2698); altitude of substratum, Bull (1955, 1957) and Roethlisberger et al. (1965). Data of central section: accumulation, Benson (1961, fig. 3, p. 22) and de Quervain (1969, fig. 32, p. 134); internal temperatures, Robin (1955, fig. 5, p. 531); altitude of substratum, Holtzscherer and Bauer (1954; fig. 13, p. 20).

In southern West Greenland, a cryptozoon-containing erratic in the Julianehåb district (Bøgvad, 1936) may have been transported by drift ice from East Greenland Precambrian areas, but Bøgvad did not exclude the possibility that the boulder was derived from unknown deposits under the ice-covered landscape of Kap Farvel.

With reference to boulders transported by the Inland Ice, they may be expected essentially to follow paths typical for boulder trains, radiating out from the ice divide. This assumption is based on evidence from glacial striae and boulder fans in formerly glaciated areas of North America and Europe, but seems to be reasonable for Greenland in view of the apparent short-distance transport of boulders to the east coast and longer transport to the west coast. Nonetheless, the precise origin and transport of boulders must be left to conjecture especially as the material may have been redeposited by pre-glacial rivers and has perhaps also followed unknown transport routes during the initial glaciation. At present, neither the available information about subglacial topography or erratics gives a clear picture of subglacial geology.

PRESENT GLACIAL CONDITIONS

Distribution of glacierized areas

In the text below distinction is made between ice-free areas, local glaciers and the Inland Ice.

Information about the areal distribution of glacierized and ice-free land is given for Greenland by Holtzscherer and Bauer (1954) as follows:

	ice-free areas.............	383,600 km^2
	local glaciers.............	76,000 km^2
	Inland Ice	1,726,400 km^2
Total area		2,186,000 km^2

Charlesworth (1957) estimated the area of the Inland Ice at approximately 1,650,000 km^2, while earlier estimates were greater. The differences must be due primarily to the varying quality of available maps, but also to the authors' delineation of different areas, especially for distinctions between local glaciers and the Inland Ice. Bauer seems to have expanded the limits of the Inland Ice to cover the regions around Kap Farvel in southernmost Greenland, the Thule area in North Greenland and in South-East Greenland, the area between Scoresby Sund and Angmagssalik. These three areas must be considered as local glaciers although they are partly confluent with the Inland Ice proper. Following this point of view, the area of the Inland Ice of Bauer, referred to above, should be reduced at least 20,000 km^2 and this amount added to the total for local glaciers.

As for estimates of its total area, ideas about the superficial form and volume of the Inland Ice have altered in accordance with available data (Mohn and Nansen, 1892; de Quervain and Mercanton, 1925; Kayser, 1928; Holtzscherer and Bauer, 1954; Bader, 1961). The total volume of glacier ice in Greenland was given by Bauer as 2,600,000 km^3; i.e., 12 percent of the world's glacier ice (Bauer, 1955a). Omitting any consideration of eventual isostatic reaction of the ocean floor, this volume represents an increase in sea level of 6 m (Bader, 1961). While local glaciers in Greenland comprise around 5 percent of the glacierized area, their volume must be less than 0.5 percent of the total ice volume in Greenland.

Climatic conditions of the present glaciation in general

As a collective expression for the parameters determining the present extent of glaciation in Greenland glaciation limits can be used. The glaciation limit is defined as the lowest possible altitude of glacier formation during the present climatic conditions as determined on maps by the summit method (Klebelsberg, 1948; Ahlmann, 1948).

The oldest map of glaciation limits in Greenland is that of Koch (1928a, fig. 90, p. 373), covering North Greenland with 250-m isoglacihypse intervals. Koch gave the lines as "snow lines". Later, in connection with the glaciological investigations around the Greenland Sea, Ahlmann (1948) compiled a isoglacihypse map of the North-East Greenland fjord zone area (also given by Charlesworth, 1957). In southern and central West Greenland, outlines of glaciation limits have been given by Weidick (1963, 1968a).

Information about the firn line of the Inland Ice has been collected by Ignat´ev (1956), Holtzscherer and Bauer (1954) and Benson (1961), and the elevations must be regarded as rather well known for the different sectors. For the firn lines of local glaciers information is more sparse, but some values are known for West Greenland (Loewe, 1934; Holland, 1961; Bull, 1963; Weidick, 1968a), East Greenland (Ahlmann, 1948), and North Greenland (Griffiths, 1960).

The glaciation limits shown in Fig. 3 are based on maps of the Danish Geodetic Institute and the U.S. Air Force. The values determined in this way seem to fit with both observed values of firn lines and the earlier maps cited above. A rise of the glaciation limits of 100 to 200 m during the period of 1920-1940 cannot be stated from these determinations, but only from evidence of isolated glaciers (Weidick, 1968a). A Canadian continuation of the isoglacihypses over Greenland can be seen from the map of Andrews and Miller (1971).

As in other mountain regions, the glaciation limits in Greenland dome up over the central parts, due to the fact that it has a more continental climate (the Massenerhebung of Klebelsberg, 1948), and at the same time the limits follow the general global rise toward the south. In Greenland, according to these two trends, the maximum glaciation limits are found in the inland area of southernmost Greenland at elevations of 1,600 to 1,800 m. The minimum values, however, are not found in northernmost Greenland, but in northern coastal areas of West and North-East Greenland, where they are around 200 to 400 m, rising to 400 to 600 m over great parts of the shores of the Arctic Basin.

Global relations of the isoglacihypses are best shown by comparison of coastal values with those of neighboring areas. A simplified map of this kind was given by Shumskii et al. (1964), where the expression used was "snow accumulation limit altitudes". The trends of these limits are nearly identical with those over the sea areas of Fig. 3.

Detailed mapping of the glaciation limits is especially of interest for comparison with the Holocene stages of local glaciations. Comparing present and ice-age climates by comparing past and present glaciation limits is necessarily speculative for Greenland because most of Greenland was fully glaciated during the ice ages.

Glacier types

According to Ahlmann (1948) glaciers can be classified by their areal distribution (morphologically), their temperature (geophysically) or their mass balance (dynamically). For the local glaciers all classifications can be used. For the Inland Ice, morphologically being only one type, its vast extent makes a physical and dynamic division necessary.

The local glaciers include examples of all morphological types described. In magnitude they vary from ice caps over 1,000 km^2, which are sometimes confluent with the Inland Ice, to cirque glaciers, piedmont glaciers and ice shelves. Piedmont glaciers have been reported from West Greenland by Milthers (1950) and Victor (1956a). Possibly the ice cover over the region south of Scoresby Sund and the Kap Farvel area must be regarded as confluent ice of transection glaciers. Glacier types in North Greenland have been listed by Koch (1928a), and his descriptions of floating glaciers in the fjords and shelf ice in Jøkelbugten (Fig. 5) are especially interesting. Occurrences of shelf ice in Greenland are apparently restricted to areas with minimum glaciation limits, but it is unknown whether or not accumulation areas occur on the shelf as described from Ellesmere Island by Lyons and Ragle (1962).

A thickness profile of one of the large local ice caps in Sukkertoppen district, West Greenland, was determined by Bull (1963). The maximum thickness was 400 m, and like Vatnajøkull in Iceland (Joset and Holtzscherer, 1954), the ice forms a lens-shaped body on top of a plateau that is slightly depressed by the ice.

With reference to dynamic classification of glaciers, studies of mass balance have been made in West Greenland where a few data are given from several localities: in the Julianehåb district (Fristrup, 1961; Weidick, 1968a), the Umanak district (Loewe, 1934; Kuhlmann, 1959), and the Sukkertoppen district (Etienne, 1940; Holland, 1961; Bull, 1963). In North Greenland glaciers in the Thule area (Griffiths, 1961) and east of Jørgen Brønlund Fjord (Fristrup, 1951; Høj, in Knuth, 1963) have been investigated, and in North-East Greenland investigations have been performed on Britannia Gletscher (Lister and Taylor, 1961), Sefstrøm Gletscher (Paterson, 1961), Bersærkerbræ (Miller, 1968) and especially on Frøya Gletscher (Ahlmann, 1942, 1948).

Nearly all the above investigations consisted of either accumulation measurements based on firn stratigraphy, or relatively short-time measurements of ablation based on stake observations.

In general terms the activity of a glacier is a function of the total input by accumulation and output by ablation occurring during a budget year. For precise definition of terms and concept of glacier activity the reader is referred to Shumskii (1950) and Meier (1962).

Fig. 5. Shelf ice. Zacharias Isstrøm, Jøkelbugten, seen from the east. In the foreground Achton Friis Ø and Schnauder Ø. Geodetic Institute route 659 A-V No. 08171 (15.08.1950). Copyright Geodetic Institute.

The scarcity and spread of mass-balance data in Greenland permit only the general conclusion that glaciers with low activity occur in North Greenland, especially in areas with high glaciation limits, while glaciers with high activity occur in South Greenland in association with low glaciation limits. Low glacier activity corresponds with slow reaction to climatic fluctuations while the increasing glacier activity toward the south and at the coast implies relatively faster response.

The mass balance has been negative for several years for most Greenland glaciers, a state which is also expressed by receding glacier fronts in West Greenland (Weidick, 1968a), North Greenland (Davies and Krinsley, 1962) and East Greenland (Sharp, 1956; Fristrup, 1961). Fluctuations of glacier lobes in general are discussed on pp. 81-93.

The Inland Ice: form, temperature and mass balance

Dimensional data of the Inland Ice are given on pp. 15 and 27. Under the present heading only information related to its physical and dynamic properties is treated.

The form of the Inland Ice

The Inland Ice is dome shaped, rising from 0 to 1,000 m a.s.l. at the coasts to around 3,000 m in the central areas. A northern dome, over 3,000 m high is separated from a southern one which is about 2,500 m high. The depression between the two domes runs from the Disko Bugt area in West Greenland to the Angmagssalik area in South-East Greenland, and has a maximum elevation of about 2,400 m a.s.l. The fact that the crest of the Inland Ice is nearer to the east coast than to the west is explained by the occurrence of higher mountain systems along the east coast (see Figs. 1 and 3) and longer westward drainage of the ice (see further p. 28).

A hypsographic curve of the Inland Ice surface was given by Meinardus (1926) and later by Bauer (1955a). From the form of this curve, Meinardus drew the conclusion that the Inland Ice was resting on a high plateau. Koch (1928a) and Demorest (1943) also arrived at this conclusion from the fact that the Inland Ice has two domes.

During the Wegener expeditions, Brockamp et al. (1933) made the first seismic investigations on the central parts of the Inland Ice, and obtained one measurement of ice thickness at the center and several near the margins. On this basis it was suggested that the Inland Ice occupied a saucer-like basin. This conclusion was questioned by Ahlmann (1941), Drygalski and Machatschek (1942) and Demorest (1943), principally on geomorphological evidence. However, subsequent seismic and gravimetric investigations by the Expéditions Polaires Françaises confirmed Brockamp's ideas (see pp. 8-9 and Fig. 3). Despite the thickness of Inland Ice, minor undulations of its surface are believed to reflect

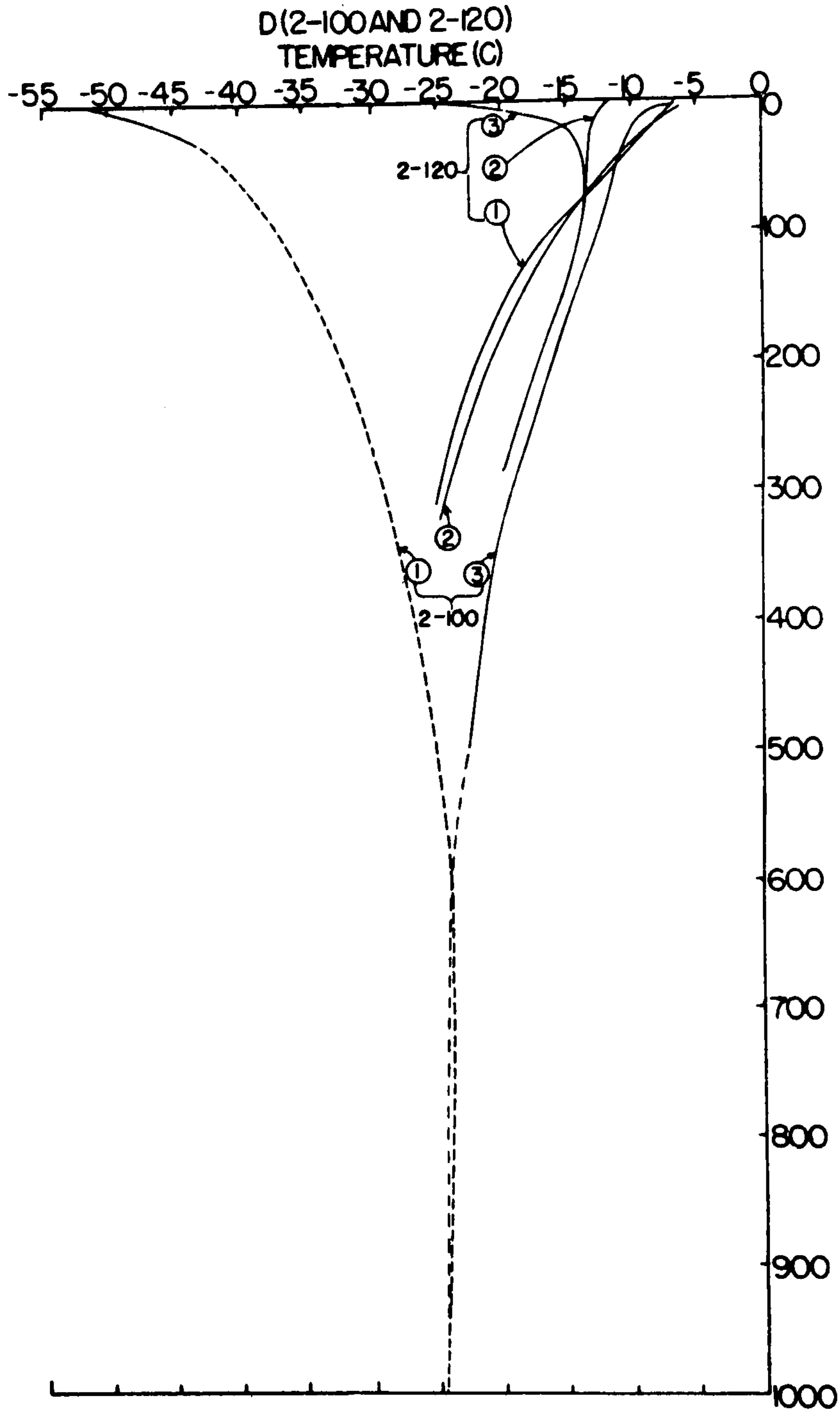


Fig. 6. Monthly temperature variations in the uppermost 10 m of the firn at Site 2 in North Greenland. Numbers for station 2-100 are: 1 = March, 2 = 7 July, 3 = 26 August; for station 2-120: 1 = 1 July, 2 = 30 August, 3 = 3 September. After Benson (1959, fig. 21, p. 25) by courtesy U.S. Army Cold Regions Research and Engineering Laboratory.

topography of the substratum (Bourgoin, 1956; Nye, 1959), but they may also be explained by differences in accumulation or by kinematic waves (Haefeli and Brandenberger, 1968).

Theoretical approximations of the form of the Inland Ice surface in simple cross-section were attempted by Mohn (Mohn and Nansen, 1892) and Meinardus (1926, 1932). However, a real basis for such calculations was first provided by laboratory investigations of the flow of polycrystalline ice (Glen, 1955). With knowledge of the rheological properties of ice, theoretical profiles were calculated for the Inland Ice and other ice sheets by Orowan (1949), Nye (1951, 1959), Vialov (1958), Weertman (1961) and Haefeli (1961). All the profiles obtained are parabolic or elliptical.

The temperatures of the Inland Ice

Where melting is negligible, snow temperature 10 m below the surface is supposedly constant within 0.3° C per year, and closely approximates the mean annual air temperature (Benson, 1962). However, the monthly variation of temperature from the surface down to 10 m is shown in Fig. 6. Temperatures at 10-m depths were mapped by Benson, and later corrected for some details by Bader (1961) and Mock and Weeks (1966). Loewe (1970) stated that mean air temperature and temperature at 10-m firn depth on the Inland Ice can deviate by up to 2° C. The information given here in Fig. 7c is essentially based on Bader's version, and it is given together with meteorological data from the coast (Blinkenberg, 1952; Lysgaard, 1969).

Monthly mean temperatures of the surface are more difficult to estimate because they require permanent observation stations. Temperature and other climatological data from the few meteorological stations in existence have been collected by Haywood and Holleyman (1961), and January and July mean temperatures are shown in Fig. 7a and 7b, respectively. The mean July temperatures are of special interest as they suggest how far onto the Inland Ice surface the summer melt extends.

As shown by annual mean temperatures, at least the southern margins of the Inland Ice surface are near 0° C and therefore must be regarded as temperate in the physical classification of Ahlmann. Temperature distribution within the body of the Inland Ice was already indicated by Rink (1853). He described water flowing from the Inland Ice margin in the Disko Bugt area in winter time, and thus showed that the ice was at melting point near the bottom even at that latitude.

Most existing knowledge about temperature within the body of the Inland Ice is based on results from deep corings (Heuberger, 1954; Hansen and Landauer, 1958; Hansen and Langway, 1966; Langway, 1967;

LEGEND

Coastal (0 m-level) mean temperatures of:

	January	July	Year
(black)	above -5° C	above +10° C	above 0° C
(horizontal lines)	-5° to -10° C	10° to 5° C	0° to -5° C
(stippled)	-10° to -20° C	below 5° C	-5° to -10° C
(white)	below -20° C		below -10° C

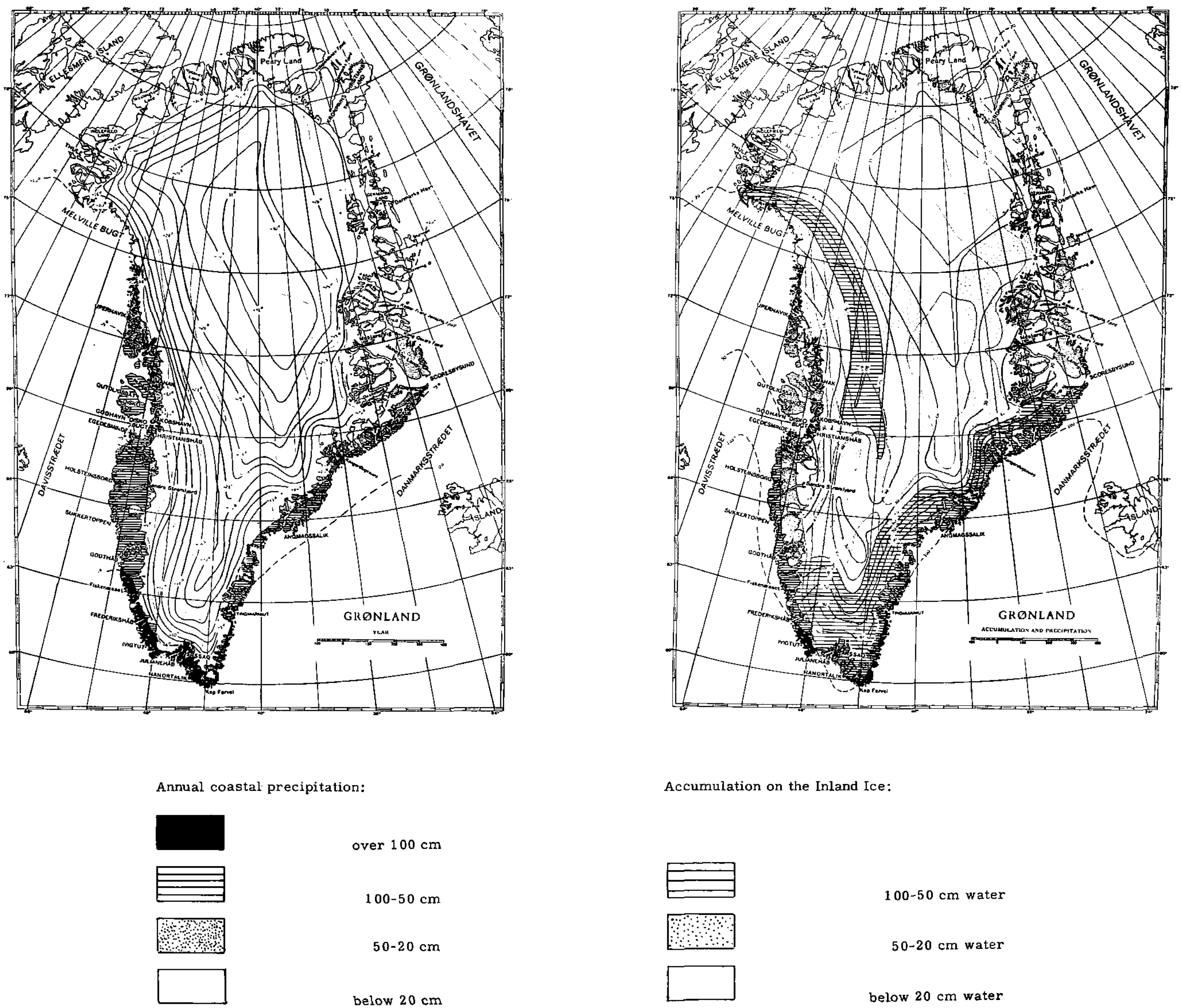


Fig. 7. a-d. Climatic factors influencing the glaciation of Greenland. The coastal stretch given according to data published by Blinkenberg (1952) and Lysgaard (1969) and on the Inland Ice according to the information of Haywood and Holleyman (1961) on temperatures and to Bader (1961) for the accumulation.

Weertman, 1968; Philberth and Federer, 1971; Philberth, 1972). A synopsis of information from the corings is given in Table 1.

Below the thermally neutral level of around 10 m, a negative temperature gradient exists for 100 to 400 m in all the boreholes. Only the Camp Century coring gives information about deeper horizons, and it shows a progressive increase of temperature with increasing depth.

Calculation of the thermal balance in profile through Camp VI and Station Centrale was made by Robin (1955). He took into consideration the cooling effect of outward and downward transport of firn from the central crest, the warming effect of thermal flux from the earth (38 $cal/cm^2/year$), and heat from internal friction between the glacier and its bed (21 $cal/cm^2/year$ for a movement of 10 m/year). Results of these calculations are shown in Fig. 4, where trends of the isotherms explain the negative gradient of the uppermost layers. Thermal conditions of the northern dome are also shown in Fig. 4b based on the temperatures measured in boreholes of Site 2 and Camp Century.

Mass balance of the Inland Ice

Determination of the total mass balance of the Inland Ice involves determination of the firn line and equilibrium line (Meier, 1962) and the loss of ice by calving, in addition to the usual measurements of accumulation and ablation.

Accumulation is known from numerous pit stratigraphies and Rammsonde readings, and annual snow accumulation throughout the last decade is well known. Summaries are given by Langway (1959), Gerdel (1961), Benson (1961), Bader (1961) and Fristrup (1966). The maps given by these authors deviate in details but are in agreement about the order of magnitude and distribution of accumulation. All show the area of maximum accumulation to be in southernmost Greenland and the minimum in North Greenland. Furthermore, they show that there is a maximum accumulation zone, trending north-south, between the west coast and the north-south trending crest of the ice surface, shown in Fig. 7d. The present mean accumulation is estimated by Loewe (1936) and Holtzscherer and Bauer (1954) to be 31 cm of water, by Benson (1961) to be 34 cm, and by Bader (1961) to be 36.7 cm. The varying estimates are esentially due to the growing amount of data; however, the distribution given in Fig. 7 must also be subject to change with climatic fluctuations. This expands the subject from the two-dimensional representation of Fig. 7d to include variations in the firn through time, as described on pp.41 and 97.

The firn line of the Inland Ice is shown in Fig. 3 as being at approximately 1,000 m a.s.l. in the north and 1,600 to 1,800 m in the south (Loewe, 1936; Bauer, 1955b; Benson, 1961). Besides discussing the firn line, Benson also modified the classification of Ahlmann by using diagenetic facies classification.

Table 1. Inland Ice corings

Site	Location	Surface m a.s.l.	Elevation of substratum, m a.s.l.	Depth of drilling, m	Surface temperature °C,	Bottom temperature °C
Camp VI	69°42'N 48°16'W	1,600	+218	125	-12.3	-16.4
Station Centrale	70°55'N 40°38'W	3,000	-181	150	-27	-27.8
Site 2	76°59'N 56°04'W	2,000	0	411	-25	-25.4
Camp Century	77°10'N 61°08'W	1,885	+495	1,390	-23	-13
Jarl-Joset	71°21'N 33°29'W	2,867	+250	1,000	-28	-30

The great firn area of the Inland Ice was divided by Benson (1961) into the following three facies:

(1) The wet-snow (formerly, soaked) facies, which becomes wet during the melting season and extends from the firn line inland to the uppermost limit of complete wetting, i.e. the wet-snow (formerly, saturation) line at which the 0° isotherm penetrates downward to melt the surface of the previous summer.

(2) The percolation facies, which is a zone of transition where melting and percolation of meltwater locally occur and cause formation of minor ice lenses and layers. It extends inland from the saturation line to a limit called the dry-snow line.

(3) The dry-snow facies, which is the zone inland from the dry snow line where only negligible melting occurs.

The facies classification furnishes a basis for interpretation of original snow accumulation, and horizontal displacements of the facies caused by secular climatic fluctuations are recorded by recognizable stratigraphic horizons.

Ablation measurements are more scarce than accumulation measurements due to the necessity of measuring the ablation at individual observation points for greater parts of the ablation season. Furthermore, surface measurements near the firn line do not accurately represent the ablation (Schytt, 1955; Nobles, 1960).

Ablation of the Inland Ice has been studied by Drygalski (1897), de Quervain and Mercanton (1925) and Loewe (1934). The mean annual ablation over the ablation area was calculated by Loewe as 110 cm of water equivalent, a figure later used by Bauer (Holtzscherer and Bauer, 1954). On the basis of Loewe's measurements, Ahlmann suggested that ablation decreases with increasing elevation of the Inland Ice surface more rapidly than on most glaciers. More recently, a mean ablation value of 107 cm was given by Benson (1962). Continuous stake measurements over longer periods were obtained by the Expédition Glaciologique Internationale au Groenland (E.G.I.G.) along a lone from Disko Bugt through Camp VI to Station Centrale (Bauer, 1960, fig. 3, p. 140; Ambach, 1963, fig. 76, p. 195), and the values recorded were of the same order of magnitude as those cited above. Maximum errors of 26 percent on individual measurements are indicated by comparing results.

The ablation area is much smaller than the accumulation area, and Bauer has given the following estimate:

Accumulation area:..	1,439,800 km^2	= 83.5 % of the Inland Ice area
Ablation area:. . .	286,600 km^2	= 16.5 % of the Inland Ice area
Total Inland Ice	1,726,400 km^2	= 100 %

When considering the mean accumulation (31-37 cm water) and ablation (107-110 cm water) over the respective areas, it must be remembered that ablation by calving is of great importance in determining the mass balance.

Estimates of mass balance have been given by Loewe (1936), Bauer (Holtzscherer and Bauer, 1954), Bader (1961) and Benson (1962). The results are as follows:

	Loewe	Bauer	Benson	Bader
Accumulation	425 km^3	446 km^3	500 km^3	630 km^3
Ablation	295 "	315 "	272 "	120-270 "
Calving	150 "	215 "	215 "	240 "

Benson's result, like Loewe's, indicates that the Inland Ice budget is nearly in balance. Differences between the figures originate in different estimates of the situation of the firn line and the magnitude of ablation, especially the calving component. Great production of calf ice is believed to occur in Disko Bugt and Umanak district on the west coast. Therefore, investigations of calving in those areas have been most intense, whereas very little is actually known about the production of other calving ice lobes, particularly those of the east coast.

A revision of the mass balance data was made by Loewe on the basis of recent information (Loewe, 1964), concluding a small gain of 158 km^3/year.

From Bauer's conclusions about the Inland Ice mass balance, he estimated that it would take 30,000 years for the ice to disappear under the present conditions, and that the Inland Ice probably could not have formed in less than 100,000 years. In contradiction to Bauer, Bader concluded that the northern part of the Inland Ice is gaining in mass, the southern part may be gaining or losing, and that the ice sheet as a whole seems to be gaining mass by accumulation in the inner parts while decreasing in total area by ablation near the margins.

The possibility of an oscillatory mechanism in the response of the Inland Ice to climatic fluctuations has been discussed by Georgi (1959), Weertman (1962) and Lliboutry (1965).

Movement and drainage of the Inland Ice

It is generally accepted that the movement of the Inland Ice is principally outward from the crest of the ice divide. The rate of movement increases from the ice divide to the firn line and then decreases to the ice margin in accordance with the flow line theory of Finsterwalder (1897). Exceptions to this general rule are given below.

Movement in the accumulation area

Movement in the firn area is very slow and mostly in a downward direction from the uppermost parts of the firn.

Both horizontal and vertical components of movement in the firn area decrease with depth. The vertical component of movement can best be given by age-depth curves, shown here as Fig. 12. The age of a particle in these curves is equal to the time elapsed since its deposition on the surface of the Inland Ice. The depth of particles of given age is shown to be dependent on local accumulation.

For Camp Century calculation of age-depth relationships attracted particular attention in relation to dating climatic variation recorded over a long time in a very deep ice core of Dansgaard _et al_. (1969, 1970, 1971). Plastic deformation of basal ice is a critical factor in such calculations (Nye, 1963; Dansgaard and Johnsen, 1969).

The horizontal component of surface movement in the firn area is difficult to measure in the absence of fixed landmarks. Methods used include:

(1) Astronomical determinations (Wallerstein, 1958; Tschaen and Bauer, 1958).

(2) Deformation of geodetic nets (Gfeller, ref. in Shumskii, 1965).

(3) Position of stations relative to relief of the substratum as determined by repeated soundings (Roethlisberger _et al_., 1965).

No conclusive results have yet been produced, but according to Bader (1961) the best estimate is that surface points move no more than tens of meters annually. For the profile through Eqip sermic - Camp VI - Station Centrale, Bauer _et al_. (1968a, fig. 7, p. 53) give a movement increasing from the equilibrium line (80 m/year) to a maximum of 160-200 m/year approximately 50 km farther east. The rate of movement then decreases to 25 m/year around 200 km from the firn line (the figures are also those used by Shumskii [1965]). The unexpected increase of movement above the firn line is explained as a perturbation caused by neighboring ice streams (fleuves de glace).

At Camp Century, horizontal velocity at the surface is as low as 3.3 m/year (Mock, in Weertman, 1968) and is uniform down to 400 m depth. Velocity then decreases gradually to zero at the bottom (Dansgaard and Johnsen, 1969, fig. 2, p. 217). Sliding at the bottom around Camp Century is therefore considered negligible. At other localities sliding of the Inland Ice along the bottom may contribute to surface movement (cf. Robin, 1955, p. 13).

Movement in the ablation area

In the ablation zone near the ice margin, surface velocities were measured at a few centimeters per day around Disko Bugt and Umanak district (Drygalski, 1897; de Quervain and Mercanton, 1925; Victor, 1956b). These results from the ice margin were verified by the E.G.I.G. expedition which also measured the previously cited increase in velocity from 18 m/year at the Inland Ice margin to 80 m/year at the firn line (Bauer et al., 1968a, fig. 7, p. 53).

Similar values were measured at the ice margin in the Thule air base area by Griffiths (1960). Near Tuto, at sites 1.0, 1.5 and 3.5 km from the ice margin, measured surface velocities were 2.0, 3.4 and 4.1 m/year respectively. Two tunnels in the ice margin at Tuto extend approximately 400 m into the ice and permit investigation of ice movements under the surface. Above the tunnels, surface velocity is 4/m year and decreases parabolically downward to 3.3 m/year at 50 m, 1.2 m/year at 75 m, and no movement at the bottom at 85 m depth (Swinzow, 1962, fig. 5, p. 223).

Carbon dioxide inclusions in ice from the Tuto tunnels have been radiocarbon dated, and ice 300 m from the tunnel portal is about 3,000 years old, while that 200 m from the portal is 5,000 years old (Oeschger et al., 1967). The age determinations and the velocities given by Swinzow led Langway (1967) to conclude that the source area of the ice was 15-25 km farther east, and that the relatively high concentrations of chemical constituents in the ice might be evidence of climatic conditions during the post-glacial climatic optimum.

At Nuna ramp (see Fig. 4), approximately 65 km northeast of Thule air base, surface velocities of 2.4 to 4.6 m/year were measured (Goldthwait, 1961), and increased inland from 8 m/year at the margin (630 m elevation) to 25 m/year (800 m). Farther inland, velocity decreases to 13-20 m/year at the firn line (1,000 m elevation) (Nobles, 1960). In contradiction to the E.G.I.G. profile mentioned above, the maximum velocity occurred below the firn line.

The drainage of the Inland Ice by ice streams

Contrasting with the above described quiet sectors of the Inland Ice are the great outlets producing calf ice. These outlets can be traced far into the Inland Ice as large, crevassed "fleuves de glace".

Rink (1857) described the ice streams and supposed the origin of the Inland Ice to have involved freezing of great rivers draining the interior. These "fleuves de glace" were later described by Bauer (1955c), and their concentration in Umanak district and Disko Bugt suggests a relationship to the possible Tertiary drainage patterns shown in Fig. 1.

The movement of the great calving lobes is characterized by block-schollen flow with most movement occurring by sliding along the bottom and little internal strain of the ice (Finsterwalder, 1950). This kind of flow must therefore be regarded as a potentially important agent in overdeepening valleys because erosive power is concentrated in parts of the glacial channel near the ice margin. However, fluctuations of the ice margin with time require that different parts of the channel are formed at various times (Weidick, 1969).

The original Tertiary drainage channels have certainly been extensively modified by subsequent intensive ice erosion. Bader (1961, p. 11) states: "An interesting question is whether the ice stream makes the depression, or vice versa; and an interesting hypothesis is that the ice stream, once started, is self-perpetuating because its ice mass, warmed up by heat of internal friction, has a lower viscosity than the surrounding ice".

For all calving outlets the highest rates of movement occur at the front. Those with the greatest velocities are the following: Eqalorutsit kangigdlit sermiat in the Julianehåb district (Jessen, 1896), Jakobshavns Isbræ and Sermeq kujatdleq in Disko Bugt (Bauer <u>et</u> <u>al</u>., 1968b), Store Gletscher, Sigssortartoq and Rinks Isbræ in Umanak district (Bauer <u>et</u> <u>al</u>., 1968b) and Upernavik Isstrøm in Upernavik district (Carlson, 1941). They all have velocities of 3 to 10 km/year.

Most other calving outlets of West Greenland have frontal velocities between 0.2 and 2 km/year as reported from West Greenland by Steenstrup (1881) and Møller (1880) for glaciers in Tasermiut and Arsuk fjords, and for glaciers in Disko Bugt and Umanak district reported by Bauer <u>et</u> <u>al</u>.(1968b).

The rate of movement of all calving outlets is not even known for West Greenland where only the main drainage basins of the Inland Ice to Disko Bugt and Umanak district have been treated in detail (Bauer <u>et</u> <u>al</u>., 1968b). For East and North Greenland the knowledge is even more sparse leaving wide latitude as to the amount of ablation by calving.

For North-East Greenland, measurements have been made on Daugaard Jensens Bræ (Olesen and Reeh, 1969) and Storstrømmen (Koch and Wegener, 1930) which had velocities of 1.8 and 4.5 km/year respectively.

In North Greenland calving outlets are believed to move relatively more slowly (Bader, 1961). A special feature in North Greenland is that calf-ice production may occur by a process of massive terminal disintegration with ice produced during several years calving on a single occasion (Koch, 1928a, Ahnert, 1963). This type of iceberg production approaches the formation of ice islands such as those described from Ellesmere Island.

An attempt to determine the velocity of the interior parts of the Inland Ice has been made by Dansgaard (1961). By radiocarbon dating the CO_2 in air bubbles of icebergs he determined the age of the glacier ice, and by measuring the proportion of O^{18}/O^{16} he approximated the elevation from which the ice was derived. From these figures, the distance and time of travel between origin and calving site could be calculated, i.e., the flow velocity. This work was done on productive lobes along the west coast of Greenland from Upernavik to Julianehåb district. Nine out of eleven icebergs were younger than 1,000 years, and the two oldest were 3,100 (Upernavik Isfjord) and 1,510 (Kangilerngata sermia) years old respectively. The distance covered by the 11 icebergs range from 60 to 460 km and the mean velocity from 110 to 270 m/year. The samples from Jakobshavns Isfjord had an age of 580 years and originated approximately 140 km from the glacier front, which indicates movement of 240 m/year (Dansgaard, 1961; Scholander et al., 1962).

Theoretical considerations of the drainage around Jakobshavns Isbræ (Bauer, 1961) indicate velocities of 5,475 m/year at front, 1,273 m/year 10 km inland and 100 m/year 100 km from the glacier front. Other theoretical estimates of drainage of Jakobshavns Isfjord were given by Haefeli (1961) and agree with Dansgaard's measurements within an order of magnitude,as do Bauer's results.

Movement of local glaciers in Greenland

Nearly all velocity measurements of local Greenland glaciers have been made in the ablation area closer to the snout than to the firn line. These measurements show a movement comparable to the inland ice margin in quiet sectors. Local glaciers in West Greenland in the Nûgssuaq peninsula, and the glaciers of Frøya and Sefstrøm in East Greenland all have velocities of 4-60 m/year (Steenstrup,1883a; Drygalski, 1897; Battle, 1952; Paterson, 1961).

Snow stratigraphy and snow metamorphism

Snow stratigraphy on local glaciers has been investigated in connection with their mass balance (see p. 19). In principle these investigations (pit diggings) do not differ from those on the Inland Ice, but their time coverage is essentially less because of greater glacier activity (cf. p. 47-19). Thus, only investigations on the Inland Ice are referred to below.

The origin of snow as a sediment starts with formation in the air around nuclei of frozen droplets or impurities. The crystals, all being hexagonal, show a great variety of form including flakes, prisms, needles, stars and cups. Both size and form are dependent on temperature of formation, the size generally increasing with the temperature.

Surface forms and deposition

Snow is mostly regarded as an eolian sediment, and dune-like surface forms and sastrugi, i.e., wind-deposited or wind-eroded irregularities, can be frequently observed. General examples are given by Shumskii (1964), and in Greenland by Benson (1962) and Bauer et al. (1968b).

According to Shumskii (1964), fresh fallen snow must be divided into two basic categories or facies:

(1) A non-eolian skeletal crystalline agglomerate, deposited under quiet conditions and relatively high temperatures, with low density, high porosity and great structural freedom of the crystals. The stratification is regular and indistinct.

(2) An eolian, solid, storm-driven snow with clastic agglomerate texture and properties opposite to those given under (1).

The two categories have many transitions and secondary alterations (see below under metamorphism) but are useful for interpretation of snow stratigraphy.

Metamorphism of snow and firn

The very complex metamorphism of snow to firn and glacier ice can be given only schematically here. For detail the reader is referred to Benson (1959, 1961, 1962) and Shumskii (1964).

In the study of Benson (1962), the Inland Ice has been treated as a "monomineralic rock formation, primarily metamorphic but with a sedimentary veneer". Benson restricted his studies to the sedimentary veneer, the maximum thickness of which he estimated to be about 90 m, but he proposed "Greenland ice sheet" as a suitable stratigraphic formation name for the whole ice mass. Thus the Inland Ice is the most widespread formation, which must be regarded as mostly Holocene in age. In facies, the formation ranges from loose sediments to completely metamorphosed ice. Benson's concept justifies a description of the Inland Ice in any general geological treatment of Greenland, but it is necessary even more because the paleoclimatological implications of investigations of the Inland Ice deeply concern aspects of Pleistocene glaciations.

However, it must also be realized that despite the likeness of ice metamorphism to other geological processes the unique physical properties of water restrict comparison to other metamorphic sediments.

Density of fresh snow depends on temperature; e.g. with the season of the year, varying from 0.01 g/cm^3 in cold, calm weather to 0.5 g/cm^3 in snow storms and wet conditions. These densities correspond with porosities of 99 to 45 percent (Shumskii, 1964). Because of their morphology, different vapor pressures will exist in different parts of individual snow crystals, and higher vapor pressures will exist in smaller crystals. There will therefore be a tendency toward rounding of all crystals and growth of larger crystals at the expense of smaller ones, resulting in 0.5 - 1.0 mm diameters of crystals.

Benson (1959) made a cumulative grain size curve of snow and firn from Greenland, comparing it with other sediments (see Fig. 9). Firn was found to have characteristics of well sorted sediments and grain sizes of medium to coarse sand.

As shown by the facies diagram of Fig. 8, further metamorphism in the soaked facies and the percolation facies takes place especially in summer, when superficial meltwater percolates downward and refreezes, forming ice layers.

Summer and winter temperature variations are also recorded by the deposition of depth hoar (Fig. 10), which is caused by rapid temperature changes in the uppermost parts of the firn. Hoar is especially indicative of autumn layers. Another process of importance is the settling of the uppermost firn layers. This firnquake (firnstoss) is described as a sudden collapse of the unstable packing of skeletal crystalline agglomerate parts of the snow (Sorge, 1935; Benson, 1959; Haefeli and Brandenberger, 1968).

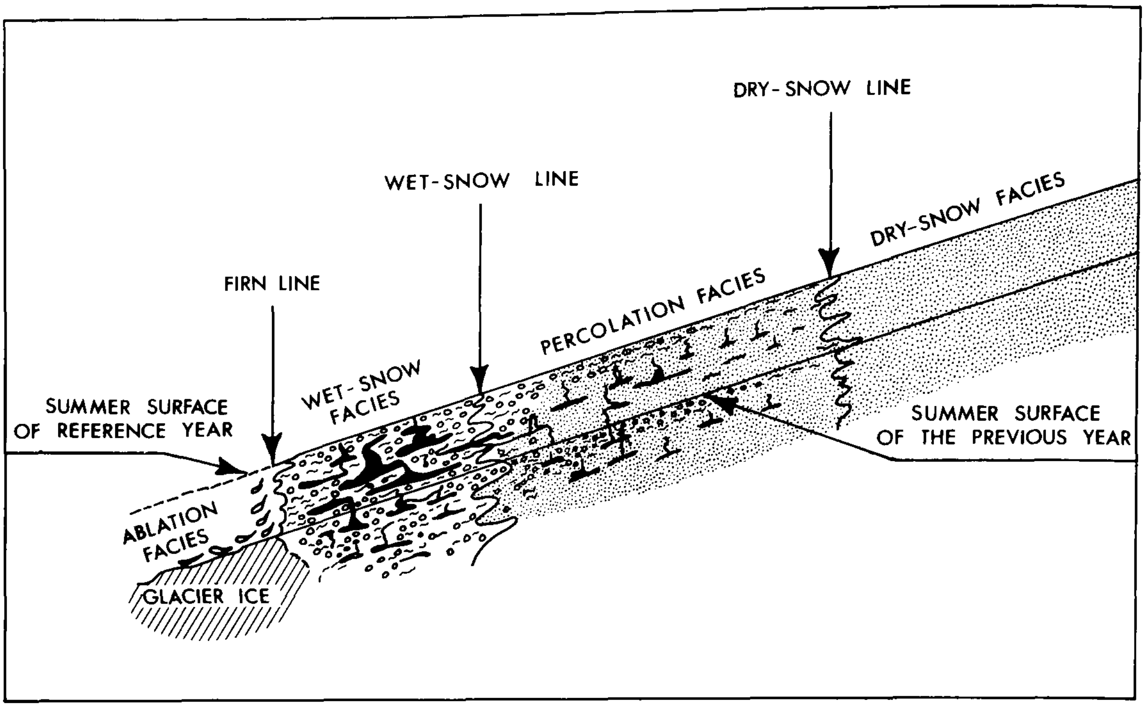
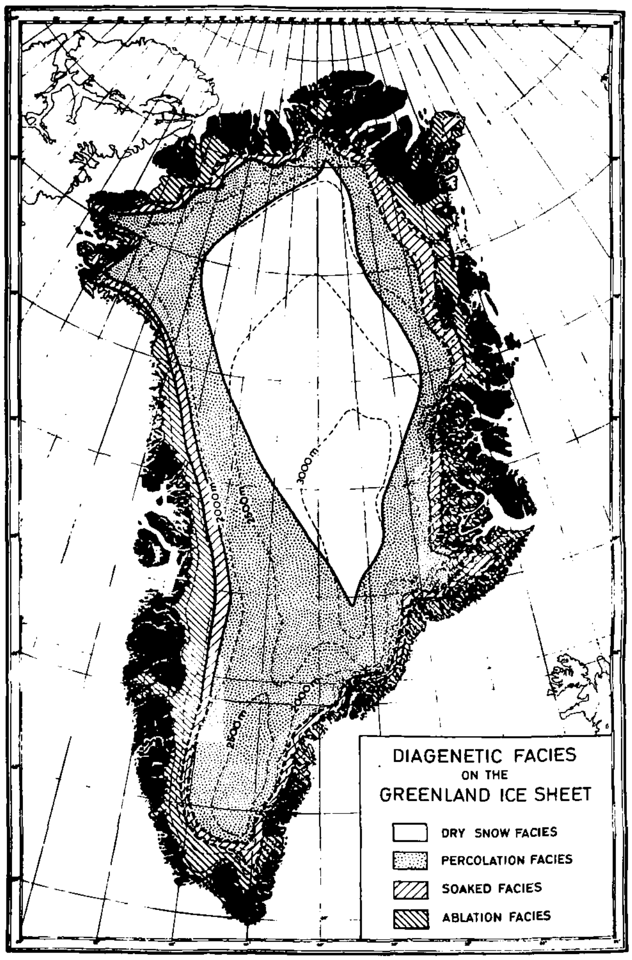


Fig. 8. Facies classification of the Inland Ice according to Benson (1959, 1961, 1962, modified 1967).
a. Schematic vertical section.
b. Extent of the facies on the Inland Ice surface.
Courtesy, U.S. Army Cold Regions Research and Engineering Laboratory.



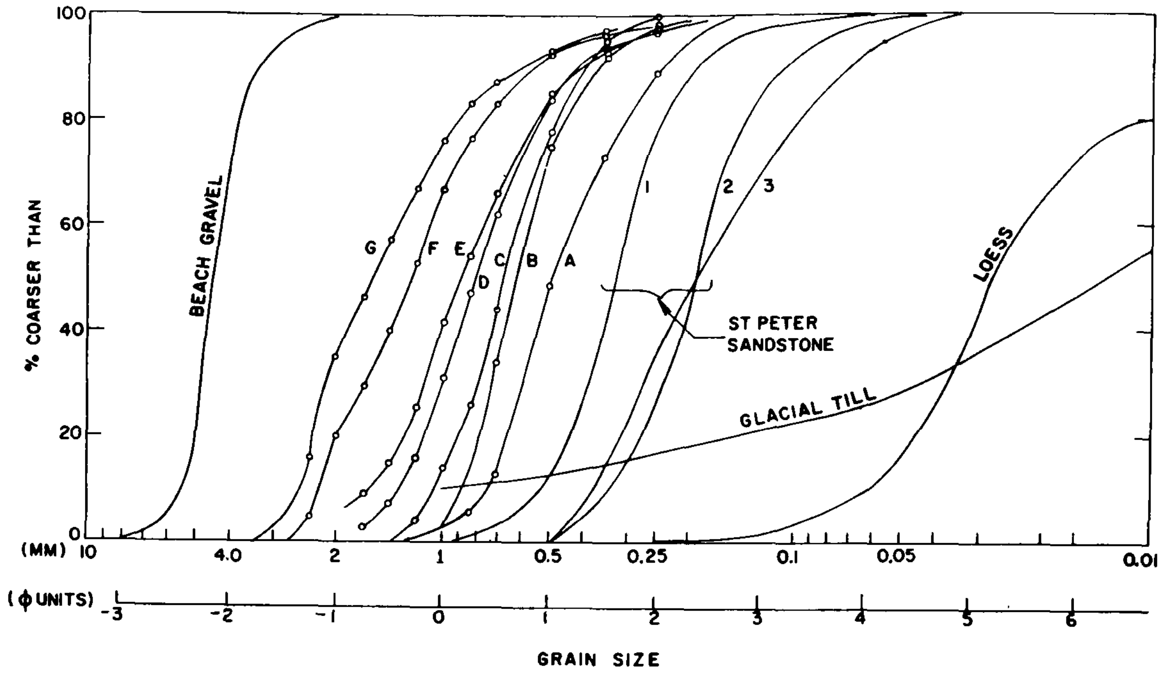
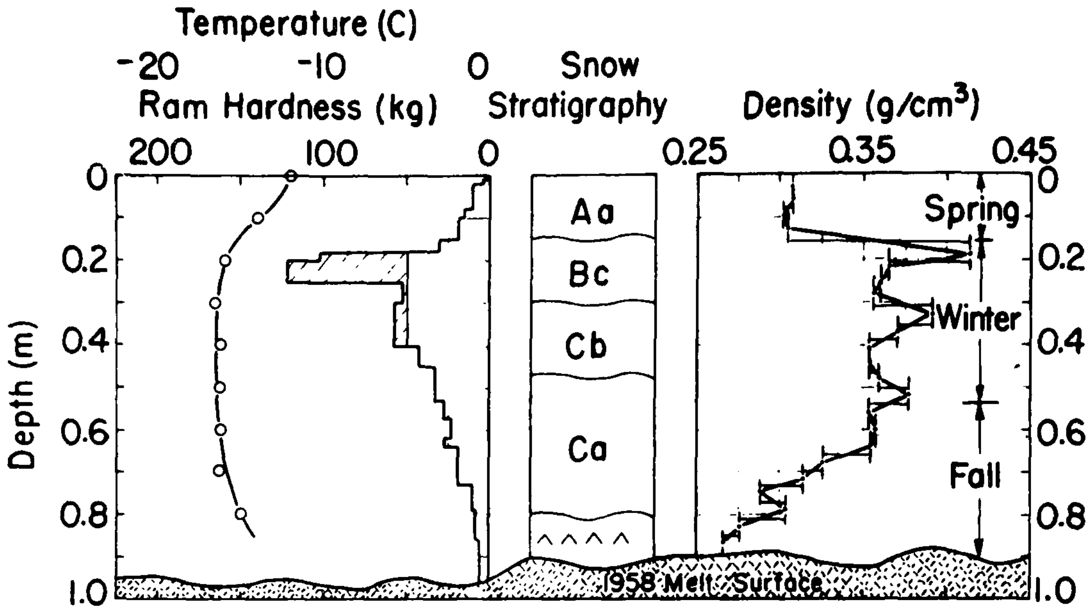


Fig. 9. Cumulative grain-size curves of Greenland snow and firn.
F-G: From layers which were wetted.
A-D: Snow layers which apparently have not been within 5° or 10° of the melting point for any significant period.
E: Loosely bonded, low density layer of slightly larger than average grain-size, including some depth hoar crystals.

The grain-size curves are compared with those of other common sediments. After Benson (1962, fig. 18, p. 28), courtesy of U.S. Army Cold Regions Research and Engineering Laboratory.



Date 5/16/59

Position 76°25'N.
67°21' W.

Elevation 788 m.

A_e, 30.2 g/cm^2

$\bar{\gamma}_p$, 0.339 g/cm^3

Relative stratigraphic symbols are as follows:

Grain Size	Hardness	
A - fine	a - soft	^ ^ ^ depth hoar
B - medium	b - medium	A_e = net accumulation
C - coarse	c - hard	γ_p = mean density for profile

Note the good agreement between the physical measurements of density and hardness and the qualitative snow stratigraphy symbolized in the center column of the figure.

Fig. 10. Plot of general observations, measurements and interpretation of data made at a surface pit study (Langway, 1967, fig. 10, p. 17). Excavation 26 km from the edge of the Inland Ice dug in the month of May. Courtesy, U.S. Army Cold Regions Research and Engineering Laboratory.

All the processes described above lead to development of a strong density gradient (Benson, 1962). Density increases from the surface down to the "critical density," which is defined as the limit between snow and firn, and has a value between 0.5 and 0.6 g/cm^3. The exact density depends on firn temperature and occurs at depths around 10 m.

At depths below the critical density, the essential processes are deformation and recrystallization of firn grains due to the increasing load of overlying firn.

The densification of firn as a function of depth is expressed in Sorge's law (Sorge, 1935; Bader, 1953), which states that the density of firn at a given depth below the surface does not change with time. Examples of depth-density curves were given by Langway (1967) and Bader (1961), and are shown here in Fig. 11.

The transition from firn to glacier ice is gradual and is limited by definition to where firn grains surrounded by channels of air space become glacier ice with air inclusions. In the example (Fig. 11), this occurs at a depth of 71 m with an average density of about 0.83 g/cm^3. In this zone of zero permeability the pressure of overlying firn is about 4.8 kg/cm^2, and the continuous increase of load with increasing depth causes a volume reduction of air inclusions so that air pressure in ice bubbles increases.

Air pressures in ice bubbles have been measured (Koch and Wegener, 1930; Langway, 1958; Scholander and Nutt, 1960), and pressures recorded up to 30 atmospheres. According to Lliboutry (1964), higher air pressures are characteristic of colder glaciers.

Lower reaches of conventional firn stratigraphy

In the present context Fig. 10 will suffice to illustrate the difference between summer and winter layers and to characterize the method of dating firn by counting annual layers. The method was developed by Koch and Wegener (1930) and Sorge (1935) and is here mentioned as "conventional" firn stratigraphy.

Benson (1962) gave examples that showed the average age of firn layers in the uppermost 10 m to be controlled by depth as well as by the local accumulation. Investigations of deeper layers are more scarce. Already in 1930 Sorge performed investigations down to 14 m depth at Eismitte [Station Centrale] (Sorge, 1935), and at this locality as well as at Station Jarl Joset investigations down to a depth of around 40 m were performed by the E.G.I.G. expedition (de Quervain, 1969). In North Greenland, firn layers have been investigated to 40 m depth at Site 2 and Camp Century. In Fig. 12 these age-depth curves are compared with those from Station Centrale and Jarl Joset, and a general dependence on accumulation is clearly seen. The curves shown in Fig. 12 are smoothed, and individual variations from these general trends reflect fluctuations in accumulation.

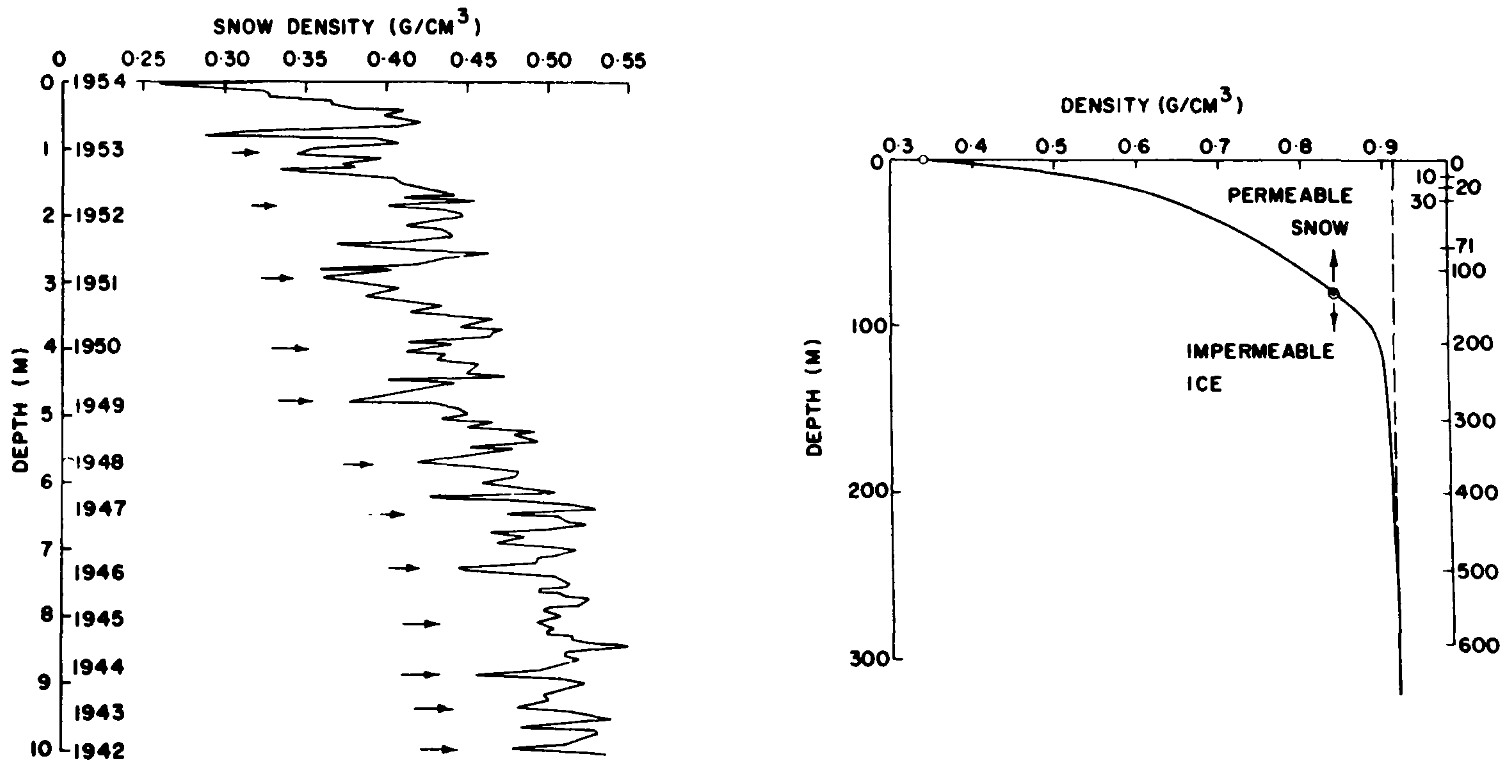


Fig. 11. Depth-density curves from the firn at Site 2, North Greenland. After Bader (1961, fig. 11 and 12, p. 17), courtesy of U.S. Army Cold Regions Research and Engineering Laboratory.

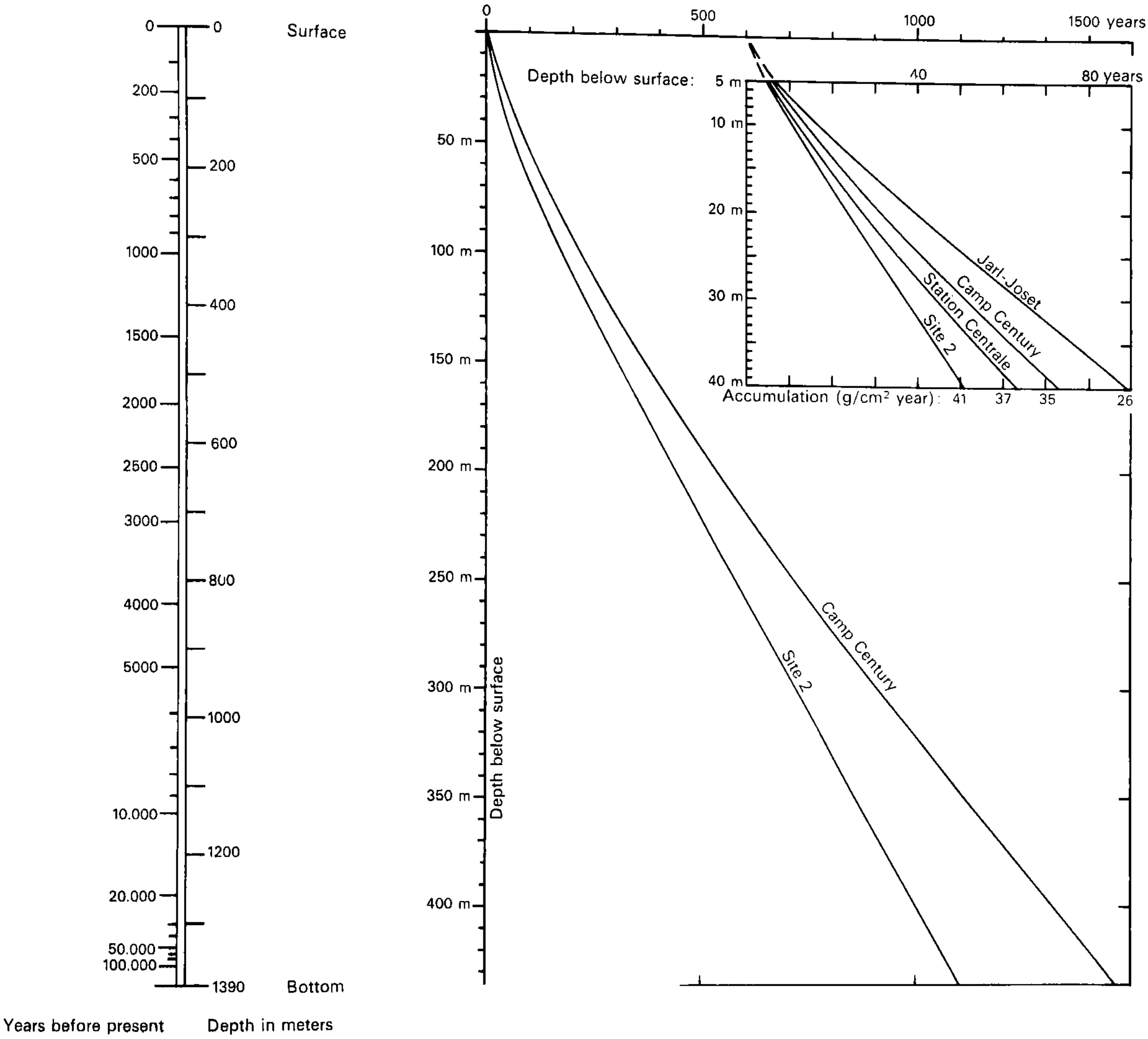


Fig. 12. Age-depth curves.
Right: Up to 40 m depths (approximately) for: Jarl-Joset, 71° 21.3' N, 33° 28.0' W, elev. 2,867 m (de Quervain, 1969); Camp Century, 77° 10' N, 61° 08' W, elev. 1,885 m (Langway, 1969); Station Centrale, 70° 54.6' N, 40° 38' W, elev. 2,964 m (de Quervain, 1969); Site 2, 76° 59' N, 56° 04' W, elev. 2,000 m (Bader *et al.*, 1955, Langway, 1969). Up to 400 m depths for: Camp Century (Dansgaard *et al.*, 1970) and Site 2 (Langway, 1969).
Left: The total extent of the Camp Century core after Dansgaard *et al.* (1969, fig. 2), given as age-depth nomograph.

For deeper strata, macroscopic investigation becomes more difficult due to thinning out and metamorphism of the firn layers. Langway concluded from investigations on the Site 2 core that it was possible to extend the conventional stratigraphic record down to around 100 m depth (Langway, 1967), and that the last structural features were clearly observed at 220 m depth. However, it seems possible to trace gross features to even greater depths. About 600 m from the bottom in the Camp Century core, B.L. Hansen found clear bubble-free ice, which was interpreted as derived from melt events 5,000 to 6,000 years ago, i.e. from the Holocene climatic optimum (Weertman, 1968). However, the core from Site 2 and the drilling log of Heuberger (1954) for Camp VI and Station Centrale seem to indicate a lower limit of structural features at only 100 to 200 m depth.

Results of conventional firn stratigraphy

The following data can be given:

(1) Melt events. Fron Site 2 Langway (1967) mentions melt events at certain levels, the deepest of which was dated by extrapolation of depth-time curves and by O^{18}/O^{16} investigations (see below). From Camp VI and Station Centrale, Heuberger reports layers of great density that may record melt events, but gives no age for the events, only their depths. However, the drillings were made in 1950 and according to the age-depth curves (Fig. 12), it seems possible to estimate a rough date of the events, at least for Station Centrale (Table 2).

Table 2. Depth-age relations at Site 2 and Station Centrale

Site 2		Station Centrale	
Depth, m	Age (surface = 1965)	Depth, m	Age (surface = 1950)
13-14	1940-41 A.D.		
25	1921 "		
43	1880-81 "	42	approx. 1885 A.D.
76-77	1824 "	71	" 1825 "
125-126	1715 "	120	" 1705 "
149-150	1660 "	129	" 1676 "

In addition to these events, there is the previously cited evidence of melt events around 5,000 to 6,000 years B.P. from the Camp Century core.

Temperature data from this century seem to indicate a close connection between climate fluctuations on the Inland Ice and those along the coastland. Gerdel (1961) compared the climatological data for Station Centrale from 1950 with those of 1931 and concluded that in this period a 1.2° mean annual temperature increase occurred. For the coastal station of Jakobshavn the increase in the same period was 1.8°.

(2) Fluctuations in precipitation. Fluctuations in precipitation indicated by the approximately 40 m of firn stratigraphy of Eismitte and Site 2 have been suggested by Bader et al. (1955), Diamond (1956, 1960) and Gerdel (1961). The values for both places agree with measured fluctuations in precipitation at Upernavik.

Date for periods previous to this century are given for Site 2 in Table 3 (Langway, 1967):

Table 3. Accumulation changes at Site 2

Site 2 (Langway, 1967)	mean accumulation, $g/cm^2/year$	
1954-57	42.3	
around 1773	34	
" 1513	37	
" 1233	41	
" 934	42	(cf. Fig. 13)

The evidence from Site 2 seems to support the idea of a dry and cold climate in medieval time, a result suggested by palynological investigations in West Greenland (see further p.97). However, it must be pointed out that recorded fluctuations of precipitation and accumulation in this century at Upernavik and Ivigtut seem to alternate in phase (Vibe, 1967), and therefore climatological data from Site 2 need not reflect fluctuations of precipitation in West Greenland's southern parts.

Foreign particles and constituents in the firn

The principal investigations on this subject have been performed on the Site 2 core, in which no micro-organisms or bacteria were found (Langway, 1962, 1967). Chemical analysis of the core has been performed for Na^+, K^+, Mg^{++}, Ca^{++}, SiO_2, Cl^-, and SO_4^{--}. Variations in concentration of single chemical constituents as well as the total amount were ascribed to meteorological changes along the coasts. For the nuclei of snow crystals at Site 2 it was found that 85 percent consisted of

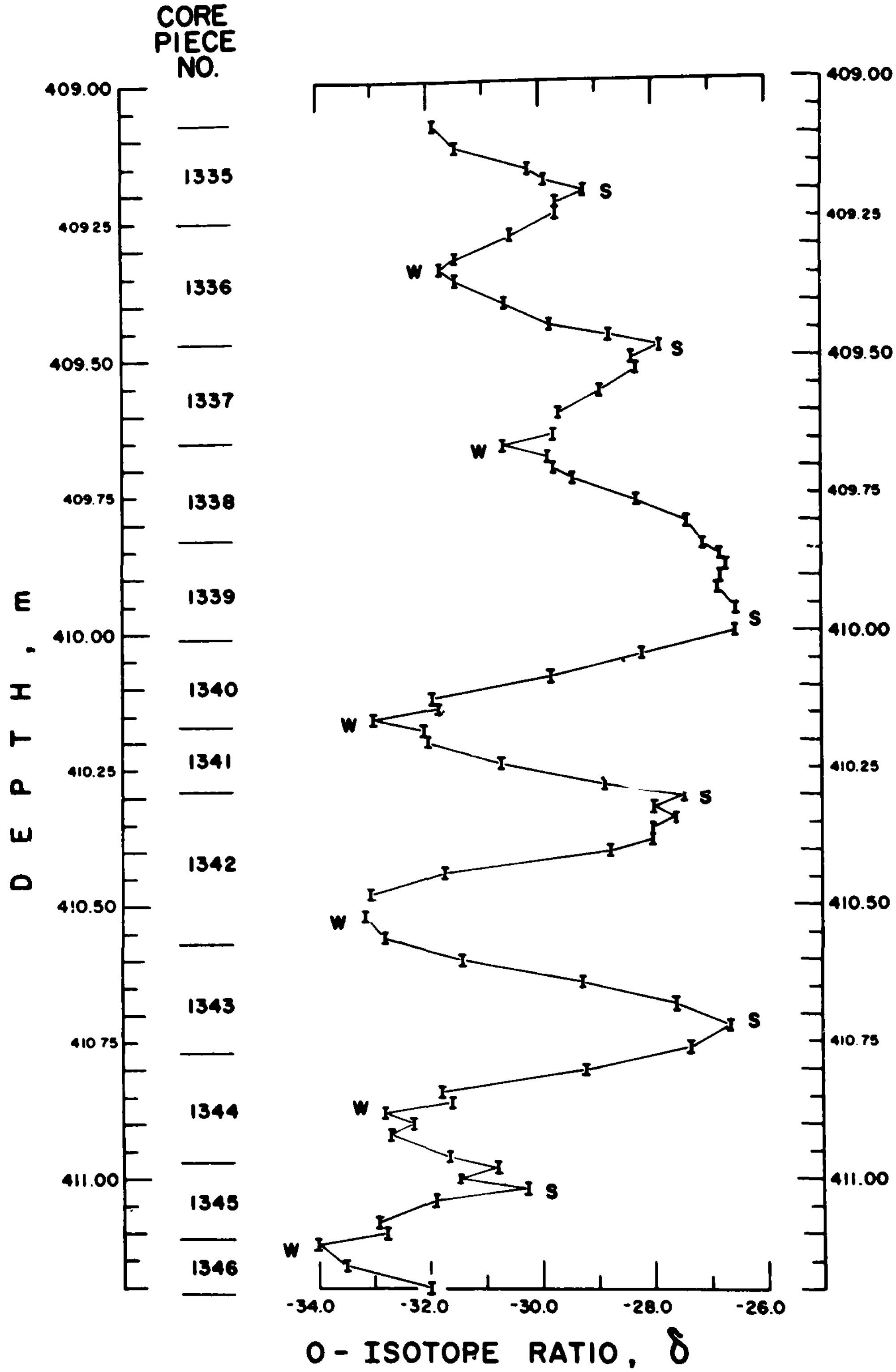


Fig. 13. O^{18}/O^{16} ratios at 411 m depth at Site 2; S = summer layers, W = winter layers. From Langway (1967, fig. 19, p. 39), courtesy of U.S. Army Cold Regions Research and Engineering Laboratory.

kaolinite, montmorillonite and illite (Kumai and Francis, 1962; Langway, 1967).

In the Site 2 core, higher annual concentrations of total dissolved constituents were found in winter snow. For long term variations a gradual decrease in total concentration from 934 to about 1400 A.D. was found, followed by a general increase to the present.

Spherules, which are droplets of tear-shaped particles with diameters of 5 to 230 μ, have also been investigated in the cores of Site 2 and Camp Century, and they seem to show a certain periodicity with depth, but until now the results only indicate another approach for stratigraphic division. Their chemical composition shows variation in metallic (mainly Fe-rich and Si-rich) content and for at least some of the Si-rich spherules a terrestrial origin was proposed. They were presumed to be produced by volcanoes or industries. However, on the basis of their chemical composition and size, Langway excluded these possibilities and concluded a cosmic origin of all spherules. It was then calculated that an average annual deposition on the earth's surface of 9×10^5 metric tons cosmic dust occurs in the form of spherules, which includes 2.1×10^5 metric tons Fe and 2.5×10^2 metric tons Ni.

Other foreign constituents involved in the stratigraphy of the Inland Ice are radioactive isotopes, which are discussed separately below.

Isotopes

In the section on movement and drainage of the Inland Ice, dating by means of isotopes was mentioned briefly. Thus far, four radioactive isotopes have been used for age dating; namely, T^3, C^{14}, Si^{32} and Pb^{210} (Dansgaard, 1967).

The basic concept in using these isotopes for dating is that they are brought to the new firn from the atmosphere at a rather constant rate and then radioactively decay at a rate suitable for dating. The amount of ice needed for obtaining a reliable date is also an important factor, which somewhat restricts isotope firn dating. The following data are taken from Dansgaard (1967) and Lorius (1963) (Table 4):

Table 4. Radioactive isotopes used for age dating of ice

Isotope	Half life, years	Dating range, (approx.) years	Amount to 1 sample
T^3	12	100	some kg
Pb^{210}	20 - 35	100	"
Si^{32}	260	1,000-1,500	1 - 5 tons
C^{14}	5,589	20,000-25,000	1 - 2 "

Added to the above is the relatively new possibility of using Ar^{39} with a half life of 269 years (Loosli and Oeschger, 1968). Investigations related to these isotopes have also been reported by Oeschger et al. (1966) and Renaud et al. (1969).

With reference to the stable isotopes of D and O^{18}, the standard isotopic composition of water is:
$H_2O^{16}/HDO^{16}/H_2O^{18}$ = 997,680/320/2,000 parts per million (ppm), and deviations from this norm are expressed by:

$$\delta = \left(\frac{\underline{a}\ \text{sample} - \underline{a}\ \text{SMOW}}{\underline{a}\ \text{SMOW}}\right) 1{,}000\ ‰$$ where a is the relative occurrence of

isotope and SMOW is Standard Mean Ocean Water. Differences between samples are essentially due to the fact that light isotopic component has a higher vapor pressure than the heavy one. δ values of precipitation have the following trends (Dansgaard, 1967):

They decrease toward the poles (latitude effect).
They decrease from coast to inland (continental effect).
They decrease with increasing elevation (altitude effect).
They decrease with decreasing temperature.

The amount required for analysis is very little (5-10 cm^3), which, together with their temperature dependence, makes the stable isotopes an important tool for discerning both annual layers and long-term climatic trends recorded in the Inland Ice.

The validity of isotopic stratigraphy is demonstrable by comparing isotope annual layers (Fig. 13) with those determined by conventional methods. The lower limit for counting annual layers with isotopes (O^{18}/O^{16}) is around 1,100-1,200 m depth in favorable cases, below which they are disturbed by diffusion. This is an important extension of conventional firn stratigraphy which only covers the uppermost 100-200 m.

In order to date layers below 1,000-1,200 m (equaling an age over 10,000 years, cf. Fig. 12) in the Camp Century core theoretical calculations of age with depth were made by Dansgaard and Johnsen (1969). The theoretical dating of the Camp Century core is shown here in Fig. 12, and it was regarded as only a first approximation by the authors. The O^{18}/O^{16} variation with depth and with calculated age yields a climate curve which shows a remarkable agreement with the trend of climatic variations known to have occurred in Europe since the last interglacial time.

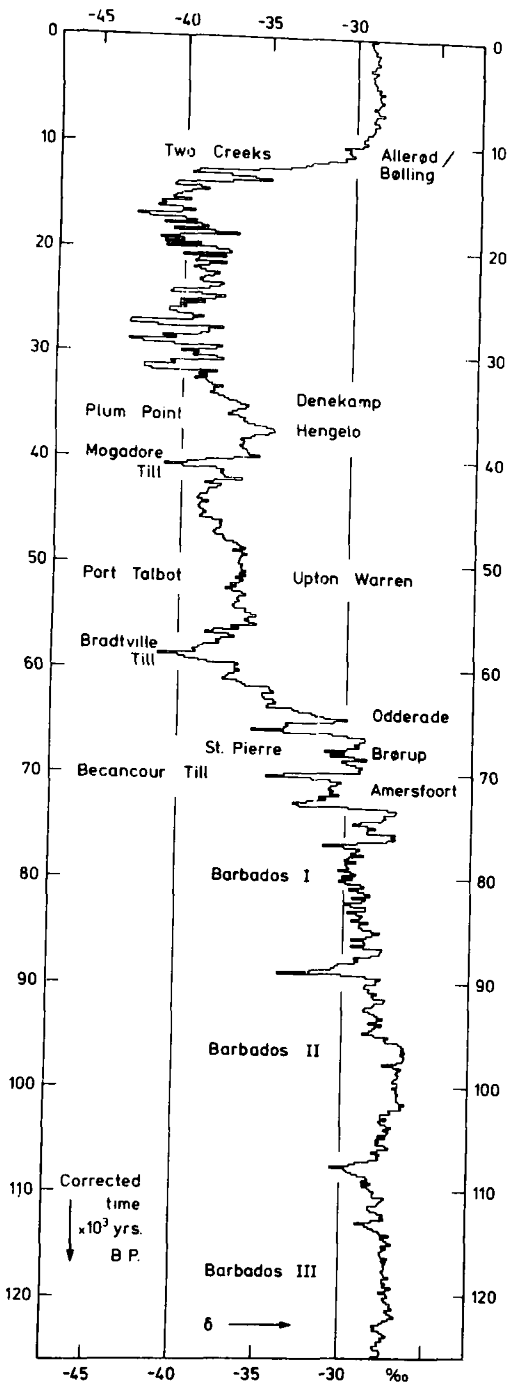


Fig. 14. The climatic curve of the Camp Century core, covering the last 120,000 years. Tentative interpretations in European and American terminology are shown to the right and to the left respectively. Horizontal scale O^{18}/O^{16} ratios. After Dansgaard et al. (1971).

For present climatic conditions on the Inland Ice it has been shown that mean temperature varies linearly with mean content of O^{18} (Dansgaard, in Renaud et al., 1969). Dansgaard, however, does not convert the δ determinations for older parts of the firn to temperature values because of unknown elevations during initial formation of deeper strata of the Inland Ice. Moreover changes may have occurred in the isotopic content of ocean water as well as in meteorological patterns and ice-flow. Therefore, the climatic curve from Camp Century, shown here in Figs. 14 and 31, gives relative warm and cold periods and relative intensity of these, but no exact temperatures.

Structural investigations in the ablation area

The problems in the classification of structures in the ablation zone were already stressed in the early investigations by Steenstrup (1883a), Drygalski (1897), Salisbury (1895) and Koch and Wegener (1930). The problems are emphasized by the variety of classifications (blue bands, dirt horizons, dirt bands, shear planes, foliation), and for the older classifications reference need only be given to the summary by Kayser (1928). Later investigations were reported in the Thule area by Bishop (1957) and Nobles (1960).

Now that stable isotopes can be used as tracers for annual layers in glacier ice (Dansgaard et al., 1960), the previous difficulties of distinguishing primary sedimentary layering from secondary metamorphic textures may be overcome.

THE PLEISTOCENE

The Pliocene-Pleistocene transition

At present no deposits are known in Greenland that with certainty can be referred to the interval between middle Tertiary and early Quaternary. A single formation, Skeldal conglomerate, near Mesters Vig in North-East Greenland, is with reservation referred to late Cenozoic - early Pleistocene (Fränkl, 1953).

Climate

The flora and fauna of Cretaceous and Tertiary age of Greenland reveal a warm climate extending into early Tertiary (Heer, 1869; Seward, 1924; Koch, 1964; Rosenkrantz, 1970). Throughout this period, however, a decrease in temperature seems to have occurred (Lowenstam, 1964; Bowen, 1966). Details of latitudinal variation in this decrease have been treated by Lowenstam (1964).

Climatic conditions throughout most of the Tertiary are not known for Greenland, and the paleotemperature curve of Fig. 15 for high latitudes is extrapolated from information for lower latitudes derived from North America and Europe (Dorf, 1955; Woldstedt, 1954; Holmes, 1965). For the early Tertiary, the curve can be checked with the information cited above for Greenland, and for the Pliocene-Pleistocene transition, with evidence from the Tjörnes deposits of northern Iceland (Einarsson *et al.*, 1967).

Tertiary glaciation of Greenland

The climatic curve of Fig. 15 must be considered as tentative but illustrates at least a general trend of Greenland climate from warm temperature conditions in the Paleogene to cold temperate conditions in the late Neogene. Neogene glaciation limits in Greenland must therefore have been comparable to the present ones over most of Europe (1,500 to 2,500 m a.s.l.), and large parts of the coastal highland may therefore have furnished centers for local glaciations. Possibly the initial glaciation already took place in the Miocene as proposed by Wager (1933). In any event, the concept of Tertiary glaciation rests on the assumption that uplift of Greenland to nearly the present elevations had already taken place.

A crucial point related to discussion of the onset of glaciation in Greenland is whether Tertiary glaciations developed into an ice sheet: The Inland Ice.

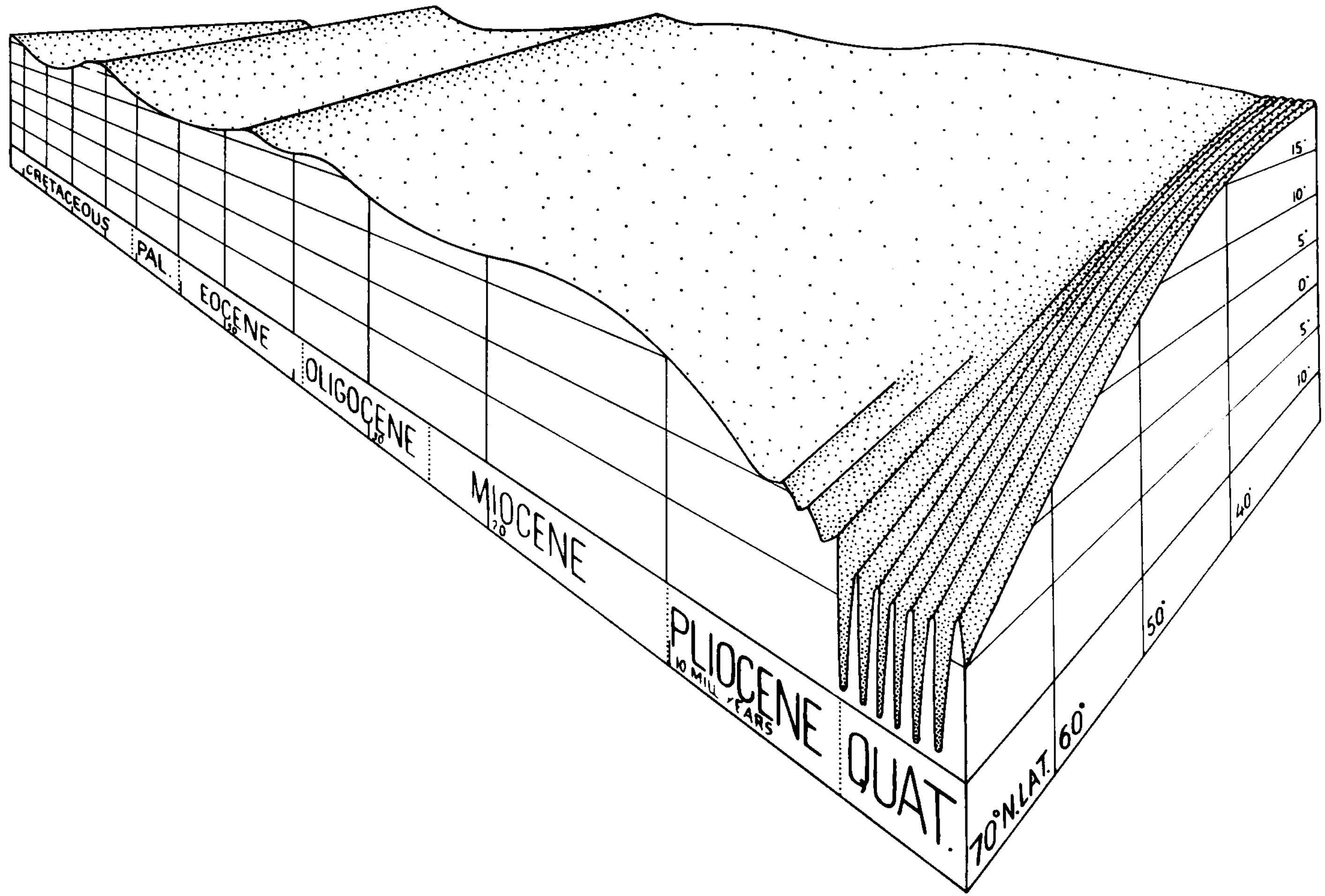


Fig. 15. Estimated mean annual temperatures for the North Atlantic region through the Tertiary and Quaternary, expressed as a function of latitude.

The autocatalysis of the Inland Ice

In contrast to the local glaciers, the Inland Ice is resting on a saucer-like substratum with a bottom near sea level (cf. Fig. 3). The Inland Ice forms its own surface topography and partly also its own climate, a process called the "self perpetuation" of the ice cover by Charlesworth (1957).

Cailleux (1952) has developed ideas about the development of the Inland Ice, originally given by Wegmann (1939), in the following way:

The initial formation on the Inland Ice is related to glaciation centers in the western and eastern marginal high mountains. With lowering of glaciation limits in the late Tertiary or early Quaternary, valley glaciers form and grow (Cailleux's phases 1 and 2). With continued expansion, the glaciers advance over the plateaus of the interior, forming great piedmont glaciers (phase 3). The piedmont glaciers then coalesce and increase in thickness because of the limited possibility of flow outward from the interior, and eventually the piedmont aprons will reach the snow line and accumulate their own firn, leading to a rapid growth of the Inland Ice (phase 4). An important factor in the rapid development of phase 4 is the decrease in absorption of the sun's radiation associated with the higher albedo of snow or ice surfaces (around 20 percent absorption by snow compared to around 80 percent by common land surfaces [Flint, 1947]).

Weertman (1962) has also given reasons for the idea of a rapidly growing ice cover. He postulated that the mechanism of ice flow indicates that a small ice cap situated in high latitudes on a continental landmass may be unstable; i.e., grow unchecked until it reaches lower latitudes and is of continental dimensions. Further, "that once an ice cap reaches this size, another instability may set in if the rate of accumulation decreases or the rate of ablation increases. The ice cap may then shrink to a small size or disappear".

Whether the Pliocene-Pleistocene boundary is defined by changes of fauna, flora, or development of ice covers, these all express the drastic climatic changes leading to the glaciations of Europe and North America. The present situation of glaciation limits shown in Fig. 3, compared to the unglaciated relief of Greenland in Fig. 1, leads to the conclusion that the Inland Ice would not develop under the present climatic conditions, but must be regarded as glacial relic, the present existence of which is due to its climatic self-perpetuation. This implies that the transition from Cailleux's phase 3 to 4 must have occurred in climatic conditions colder than at present. It is therefore believable that initial development of the Inland Ice proper was contemporaneous with the onset of glaciations of North America and Europe; i.e., at the beginning of the Quaternary.

The glaciation of Greenland in interglacial times

A total deglaciation of Greenland has been supposed but is incapable of direct proof (Flint, 1947; Charlesworth, 1957). The slight rise in sea level of 5 to 10 m which would result from a total melting of the Inland Ice cannot be detected by evaluating the fluctuations of interglacial sea levels. However, the Inland Ice existed at least throughout the last interglacial (Sangamon-Eem) in North Greenland, as demonstrated by the presence of interglacial ice layers in the Camp Century core (see p. 44).

The Inland Ice does seem to have been smaller than at present during some parts of interglacial times. This may be concluded from concretions of marine sediments brought to the Inland Ice margin from its substratum at different localities in West Greenland, from Julianehåb to Upernavik areas (Kornerup, 1879; Gripp, 1932). Some of the concretions are possibly of Holocene origin, but Bryan (1954) has demonstrated the probability that some are also of interglacial age by identifying *Picea mariana* pollen in two of the concretions. In a sample from Godthåbsfjord, the quantities of pollen were sufficient to imply its growth in the area, but it was less certain for one from Frederikshåb district. *Picea mariana* has not been present in Greenland in Holocene time.

Some radiocarbon ages of marine shells and drift wood are greater than 32,000-35,000 years, and their origin must therefore be regarded as possibly interglacial. The information in Table 5 has been obtained in this respect (cf. Figs. 16 and 17):

Table 5. Reported interglacial occurrences in Greenland

Locality	Dating	Reference
Nordre Isortoq, West Greenland	K-1555	M.Kelly (oral communication)
Nugssuaq peninsula, "	K-1545	Rosenkrantz (1968)
Saunders Ø, Thule area, North Greenland	W-74 W-75	Davies *et al*. (1963)
Øvre Midsommersø, Peary Land, North Greenland	K-1445 K-1447	Fredskild (1969a)
Mudderbugt, Peary Land, North Greenland	K-314	Trautman and Willis (1966)

While the first four samples are shell material, the last ones from Peary Land are all wood imbedded in marine beaches (Midsommersøer) or in what are believed to be Wisconsin moraines in the easternmost parts of Peary Land.

Fig. 16. North coast of Nûgssuaq peninsula, Umanak district, West Greenland, seen from the north. The picture shows the stretch between the rivers Sarfâgfîp kugssinerssua (left) to Aorrussap kugssinerssua (right). Coastal central part is the Patorfik locality and behind this the mountain of Qilertínguit (elev. 1,977 m), both mentioned in the text. Geodetic Institute's route 514 K-SV no. 29 (15.07.1948). Copyright Geodetic Institute.

Fig. 17. Slusen, Peary Land, North Greenland. Slusen, a passage connecting the lakes of Øvre Midsommersø (left) with Nedre Midsommersø (right), is mainly formed as a lake- and kame terrace system. Geodetic Institute's route 548 C-N no. 4266 (15.07.1950). Copyright Geodetic Institute.

Caves of interglacial age were reported by Davies and Krinsley (1960) from the Centrum Sø area in North-East Greenland, and a possible interglacial peat deposit situated between overlying "schotter" and underlying ground moraine was described from Loch Fyne farther south in North-East Greenland by Backlund (1931b). In West Greenland two separate profiles from Julianehåb district (Weidick, 1963) and from Holsteinsborg district (de Quervain and Mercanton, 1925) show moraines separated by fluviatile or marine deposits. For these areas and Loch Fyne, the whole series might represent an interstadial sequence.

Greenland during the glacial ages

Multiple glaciations of the Greenland coastland

As a consequence of the acceptance of interglacial deglaciation of parts of Greenland, the existence of earlier glaciations of the Greenland coasts must be accepted.

Double cirques and U-valleys have been taken as evidence for a two-fold glaciation on the west coast by de Quervain and Mercanton (1925) and Belknap (1941), and on the east coast by Flint (1948). However, the erosion cycles necessary for such forms may result from fluctuations of a single glaciation.

In general, the formation of large scale features such as cirques and especially fjords, U-valleys, and strandflats may be due to repeated glaciations, and their occurrence along the Greenland coast therefore suggests former glaciations.

Rock basins and cirques

Rock basins and cirques are features for which a glacial origin is beyond doubt, and both are common over all of Greenland. With reference to cirques, Boyé (1950) and Graff Petersen (1952) concluded from their studies in West Greenland that cirque formation is characteristic of the initial stages of glacial erosion, and that glacial striae on the sides of cirques in coastal stretches showed they had been overridden by continental ice at a later stage.

Elevations of cirque thresholds give an approximate value of the glaciation limit during the coldest periods, which triggered subsequent continental glaciations of the coast. They may also indicate the approximate height of the glaciation limit during the glaciations (see also p. 64).

Fjords

Fjords occur along all the coasts of Greenland. The maximum lengths of fjords on the west coast occur in middle West Greenland, where they range between 130 and 180 km. The fjords, like those of Norway, consist of a series of deeper basins separated by shallower thresholds. Most fjords in West Greenland and East Greenland terminate as troughs in the shelf along the shore and do not continue to the edge of the shelf (cf. p. 64).

The greatest depths measured in West Greenland fjords are 1,123 m in Kangerdluk fjord, Umanak district (Sorge, 1932), 1,055 m in Upernavik Isfjord (Birket-Smith, 1928), and 920 m in Disko Bugt (Kongelige Søkort Arkiv, Chart 1500). In general, maximum depths of most larger fjords are 600 m.

In East Greenland, the fjords differ in some degree from those of West Greenland. Many of them continue as depressions across the shelf, and the land surrounding the fjords is often higher than in West Greenland. Also, in the East Greenland areas of sedimentary rock, the fjords are wider and not as steep-sided as in areas with metamorphic rocks. The fjords reach maximum development in the great complex of Scoresby Sund and Kejser Franz Josephs Fjord. Scoresby Sund fjord complex (Fig. 18), with length of around 300 km, is one of the deepest fjords in the world. The maximum depths in East Greenland fjords seem to average around 400 m, except in the crystalline rocks of the East Greenland Caledonian belt, where greater than 800 m depressions have been measured in Kempe Fjord (LeRoy, 1948), and 1,450 m in Scoresby Sund (Thorson and Ussing, 1934).

With reference to the origin of the fjords, pre-glacial erosion during upheaval of the land, periglacial processes, and glacial erosion all have to be considered agents of importance. In West Greenland, fjords are considered to have at least a partly tectonic origin (Ussing, 1912; Krueger, 1928; Wegmann, 1938). In North Greenland, a similar tectonic origin is cited by Koch (1928b) for the area between Melville Bugt and Smith Sound. Farther north, from Robeson Channel to Peary Land, fjords have the same relationship to structure as the fjords of East Greenland, and may have a similar origin. However, conflicting opinions about the latter have been given. Ahlmann (1941) pointed out that fjords cut across the major structural elements and probably owe their origin to erosion, while Backlund (1931a) and Frebold (1932) stress the influence of tectonics.

At least the final form of the fjords is due to glacial over-deepening, especially by ice streams (see also p. 30). Greatest erosional power is confined to the marginal areas of the Inland Ice,

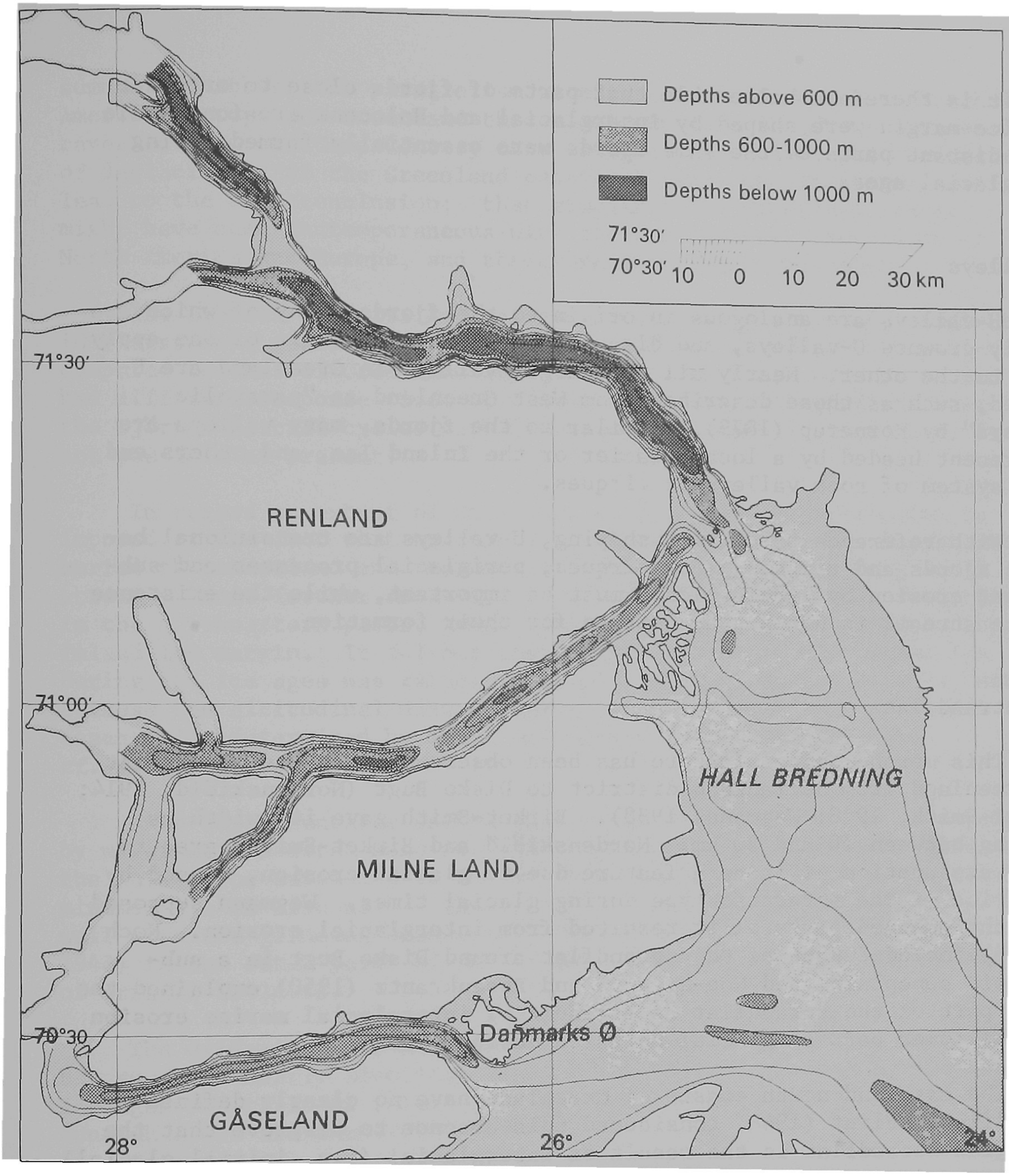


Fig. 18. Depths of inner part of Scoresby Sund. Data compiled from Thorson and Ussing (1934) and Chart 2600 (1:400,000) of the Royal Hydrographic Office.

and it is therefore believable that parts of fjords close to or under the ice margin were shaped by interglacial and Holocene erosion, while more distant parts of the same fjords were essentially formed during the glacial ages.

U-valleys

U-valleys are analogous in origin to the fjords, many of which are simply drowned U-valleys, and discussions about the origin of one apply also to the other. Nearly all the larger valleys in Greenland are U-shaped, such as those described from West Greenland as "parabolic valleys" by Kornerup (1879). Similar to the fjords, many valleys are at present headed by a local glacier or the Inland Ice, and others end in a system of rock valleys or cirques.

With reference to glacial shaping, U-valleys are transitional between fjords and cirques. For cirques, periglacial processes and subsequent erosion by glacier lobes must be important, while the existence of ice streams is not a prerequisite for their formation.

The strandflat

This morphological feature has been observed along the west coast of Greenland from Julianehåb district to Disko Bugt (Nordenskiöld, 1914; Birket-Smith, 1928; Wegmann, 1938). Birket-Smith gave its width as ranging between 10 and 30 km. Nordenskiöld and Birket-Smith agree in their explanation of it as a feature due to glacial erosion, caused by shelf ice of the Antarctic type during glacial times. Wegmann supposed that the emerged parts of it resulted from interglacial erosion. Koch (1928b) included part of the strandflat around Disko Bugt in a sub-Mesozoic peneplain, and Noe-Nygaard and Rosenkrantz (1950) explained the upper part of the strandflat as a result of interglacial marine erosion and the lower part as glacial abrasion during glacial ages.

The east and north coasts of Greenland have no clearly defined strandflat. Flint (1948) considered this absence to disprove that the Norwegian strandflat is the result of interglacial (i.e. initial glacial) marine erosion, and he supported the suggestion of Ahlmann (1919) that the strandflat is a downwarped terrestrial erosion surface. The various hypotheses about this feature were summarized by Birket-Smith (1928) and Werenskiöld (1951).

Number and extent of glaciations of the Greenland coastland

Multiple glaciation of the Greenland coast has only locally been demonstrated *in situ* in East Greenland. Nonetheless, the evidence summarized above, the parallelism of climatic fluctuations in North America, Europe and Greenland throughout the last 120,000 years as revealed from the Camp Century core (see pp.44 and 50), and the history of deglaciation of the Greenland coastland since the Wisconsin, all lead to the same conclusion: that glaciation of Greenland coasts might have been contemporaneous with the continental glaciations of North America and Europe, and therefore occurred repeatedly.

The Tjörnes deposits in Iceland (Einarsson et al., 1967) indicate 10 Pleistocene glacial cycles, but the climatic situation of Iceland compared to that of Greenland may indicate that Icelandic ice covers had different response times to climatic changes. A correlation of the Tjörnes glacial cycles with Greenland continental glaciations is not possible at present.

In regard to extent of Greenland glaciations, a restriction is imposed by the fact that Greenland is an island: once the Inland Ice buried the coastland and reached the sea, calving and melting would strongly limit further expansion. At some places, however, especially in the northeastern parts, great ice shelves were developed from the Inland Ice margin. It follows that if the extent of the Inland Ice during the ice ages was rather constant, then so was its surface form, because the altitudinal distributions of continental glaciers are essentially determined by their horizontal extent (Robin, 1964: cf. p. 21).

Some high mountains on the west coast of Greenland are surrounded by moraines ("nunatak moraines"); their height indicates a profile of the Inland Ice extending to the shelf off Greenland which suggests a minimum age of Wisconsin. The elevations of these moraines fit fairly well with the grinding boundaries ("Schliffgrenzen" of Klebelsberg, 1948) in the outer parts of some fjords and with the uppermost limit of erratic boulders.

The evidence given above is still too scattered to be conclusive but suggests nearly identical forms of the Inland Ice during different ice ages. Details of Inland Ice form and the existence of potential nunatak areas are therefore treated with the Wisconsin ice cover below.

In East Greenland, Funder and Hjort (1973) have produced strong evidence for an ultimate limit of the youngest Inland Ice glaciation leaving the eastern parts of the Geographical Society Ø and Traill Ø as well as most of Jameson Land ice-free. Liverpool Land was presumably partly covered by local glaciers. This youngest glaciation is correlated to younger Wisconsin, whereas an older continental glaciation seems to have covered the outer coast.

For North Greenland, Davies (1972) advocates the thought that due to increasing aridity throughout the Quaternary, the Inland Ice reached its greatest extent during possibly Mindel (Kansan), whereas Riss (Illinoian) was less extensive and Würm (Wisconsin) had an extent not much greater than at present. However, this idea seems to conflict with the fact that the late and fast Holocene uplift of North Greenland (see further p. 73) involves the whole of North Greenland and can only be explained as a reaction on a young, presumably Wisconsin, glaciation.

THE MAXIMUM GLACIATION OF GREENLAND

As shown in the previous section, knowledge of the Pleistocene glaciations is still too scattered to allow a delineation of the Inland Ice extent during glacial times. In the light of the information from East Greenland it seems feasible that the Wisconsin extent of the Inland Ice was smaller than the maximum extent. However, since a widespread Holocene uplift of land took place, the Wisconsin ice cover must also have covered large parts of the Greenland coastland.

Surface conditions

West Greenland

In the southernmost parts, a nearly total glaciation of the outer coastland, extending to the banks off Julianehåb district, was presumed by Jessen (1896) on the basis of erratic boulders and glacial striae. This agreed with observations by Steenstrup (1881) that 700 to 800 m was the limit of glacial striae in Kitdlavat east of Julianehåb. Accordingly great parts of the alpine highland must have been nunataks.

Farther north, in the Frederikshåb district, observations around Arsuk also showed the upper limit of former ice cover to be around 700 m (Bendixen, 1921; Upton, personal communication). North of Frederikshåb district an increase of the former ice cover surface is suggested by observations that the limit is about 950 m in southern Godthåb district and 1,250 m in northern Godthåb district (Bendixen, 1921). However, these last limits occur relatively farther inland.

In general, it seems that the highest western coastal areas of most of West Greenland have upper limits of former ice cover of 600 to 800 m a.s.l., such as those reported from Holsteinsborg and Egedesminde districts (Kornerup, 1881; Pjetursson, 1898; Kelly, 1969) and from the island of Disko, Umanak and Upernavik districts (Steenstrup, 1883b; Ryder, 1889; Henderson, personal communication). A deviation from this trend is a limit of former ice cover around 900 m a.s.l. near the head of Disko Nordfjord (Steenstrup, 1883b) which may be due to the influence of the Disko highland on the surface topography of the ice cover. Another exceptional value of only 300 m was reported by Pjetursson (1898) from the island of Rifkol south of Egedesminde. The low value may be explained by the influence of relatively flat substrate topography on the ice cover, similar to the present Inland Ice margin surface in Holsteinsborg and Egedesminde districts.

For Holsteinsborg-Egedesminde and Umanak districts the above values are based on nunatak moraines, upper limits of erratic boulders or grinding limits. An eastern continuation of the former ice surface can be followed in the Umanak district, where the upper limit of erratics on Ubekendt Ejland and near Ikorfat is 900 to 1,100 m, and about 1,500 m at Qilertinguit and in Uvkusigssat fjord. This surface can be traced to the western parts of Nugssuaq by nunatak moraines 600 to 800 m a.s.l.

In Melville Bugt, erratic boulders on the top of Inugsulik island near Djævelens Tommelfinger ("Devils Thumb") (Fig. 19) up to 550 m a.s.l. witness an extent of the ice cover over wide parts of the shelf (Ryder, 1889).

North Greenland

The absence of ice cover in the northern parts of Peary Land has been stressed in current literature (e.g. Charlesworth, 1957, p. 726). The uniformitarian argument that low precipitation of the area means it was partly ice-free during the ice ages may be questioned when present precipitation is compared with the wide spread of local glaciers and the glaciation limits shown in Figs. 2 and 3. Recent investigations (Davies, 1963) reveal that the Inland Ice during the Wisconsin buffered against local ice caps of the high mountains in northern Peary Land, and the local ice covers extended northward into the Arctic Basin as shown by the moraines of Kaffeklubben Ø.

In North Greenland, Koch (1928a) presumes that isolated mountains such as Djaevelens Tommelfinger, Conical Rock and Dalrymple Rock ("Dalrymple Island") may have been isolated nunataks during the glaciation. Erosional forms alone, however, are not always indicative of nunataks. The northern extent of the Ellesmere Glacier Complex is unsure, but at least Davies and Smith (in Smith, 1961) show a western extension of the Inland Ice to around Alert in the northern part of Ellesmere Island, a fact which fits fairly well with the thickness of the Inland Ice indicated by erratic boulders on Beaumont Ø and Dragon Pynt in North Greenland, respectively 450 and 1,000 m a.s.l. (Koch, 1928a).

From Alert and southward, the limit between the Inland Ice and the Ellesmere Glacier Complex was given by Christie (1967), who indicated that the Ellesmere coastland was covered by the Inland Ice as far south as Kennedy Channel, and south of Kennedy Channel the Canadian and Greenland ice covers merged over what is now international water. The two ice covers were confluent until somewhere west of Carey Øer. On these islands, marks of glaciation (Bendix-Almgreen et al., 1967) indicated an ice cover from Greenland which at its maximum was more than 300 m thick.

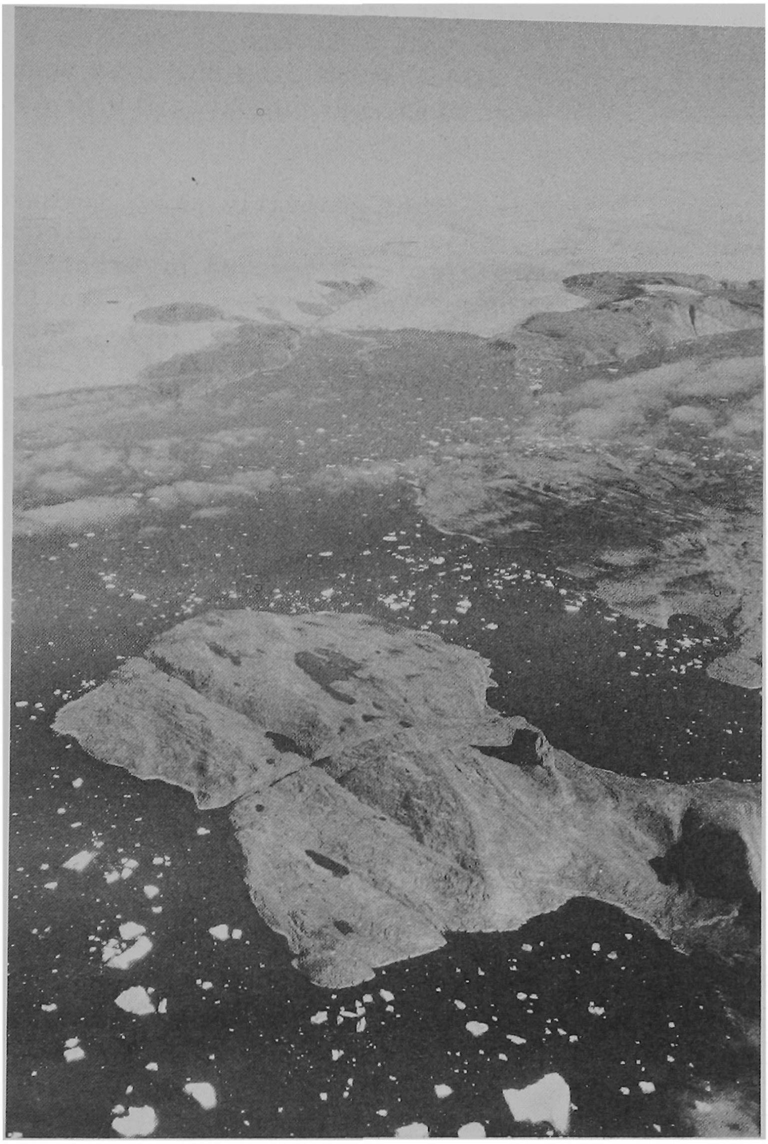


Fig. 19. Devils Thumb, Upernavik district, West Greenland, seen from the west. In the background the Inland Ice margin. Geodetic Institute's route 535 A-Ø, no. 3969 (02.08.1949). Copyright Geodetic Institute.

East Greenland

Data from the outer coast of East Greenland indicate an upper limit of erratics 400 to 800 m a.s.l. at Hold with Hope, Clavering Ø, and Jackson Ø (Bretz, 1935). These values agree with the idea that the east end of Traill Ø was a nunatak more than approximately 700 m a.s.l. (Noe-Nygaard, 1932).

The surface of the former ice cover generally seems to have had a steeper gradient than in West Greenland. Peaks between the outer coast and central coastland of East Greenland are covered by erratics to elevations of 1,400 to 1,800 m on Hudson Land, Gauss Halvø, Traill Ø and Ymer Ø (Bretz, 1935; Noe-Nygaard, 1932; Backlund, 1931b). Only south of Kejser Franz Josephs Fjord does a less steep surface of the former ice cover seem to be indicated by an upper limit of 500 to 800 m around Mesters Vig (Pessl, 1962; Washburn, 1965; Cruickshank and Colhoun, 1965; Lasca, 1969). A maximum extent of the ice cover is here at least older than younger Würm (Wisconsin) (see p. 57).

Extent of the ice cover

In West Greenland a limit of ice cover is indicated by the banks (cf. Fig. 20). Detailed hydrographic charts show several closely spaced, N-S trending ridges on the surface of the banks, indicating marginal moraines from the ice ages. Their age is not known with certainty, and they are regarded here only as a zone. An eustatic lowering of about 100 m during the Wisconsin means that great parts of the ice margin over the banks have rested on dry land, while the intervening deep troughs were presumably occupied by ice streams.

The northern confluence of the Inland Ice with the Canadian Ellesmere Glacier Complex has been described above. While restricted to the west by Ellesmere Glacier Complex, the eastern part of the northern edge of the Inland Ice terminated in the local glaciations of Peary Land. Thus, the Inland Ice proper reached the Arctic Ocean only along a border about 200 km long.

The border between the Inland Ice and the high mountains of northern Peary Land was first defined by Koch (1928a, b) on the basis of a northern limit of erratics. This line was later altered slightly by Troelsen (1952a). In the easternmost parts of Peary Land, the extent of the Inland Ice is shown by moraines. Dating of wood near Mudderbugt (see p. 50) gave an age of more than 32,000 years B.P. The very complex form of these Wisconsin moraines, as most recently outlined by Davies (1963, fig. 3), seems to indicate a seaward expansion in the eastern parts of Peary Land as shelf ice. This should also explain the occurrence of erratic boulders, possibly transported by the Inland Ice, found along the northern shores of Peary Land (Dawes, 1970).

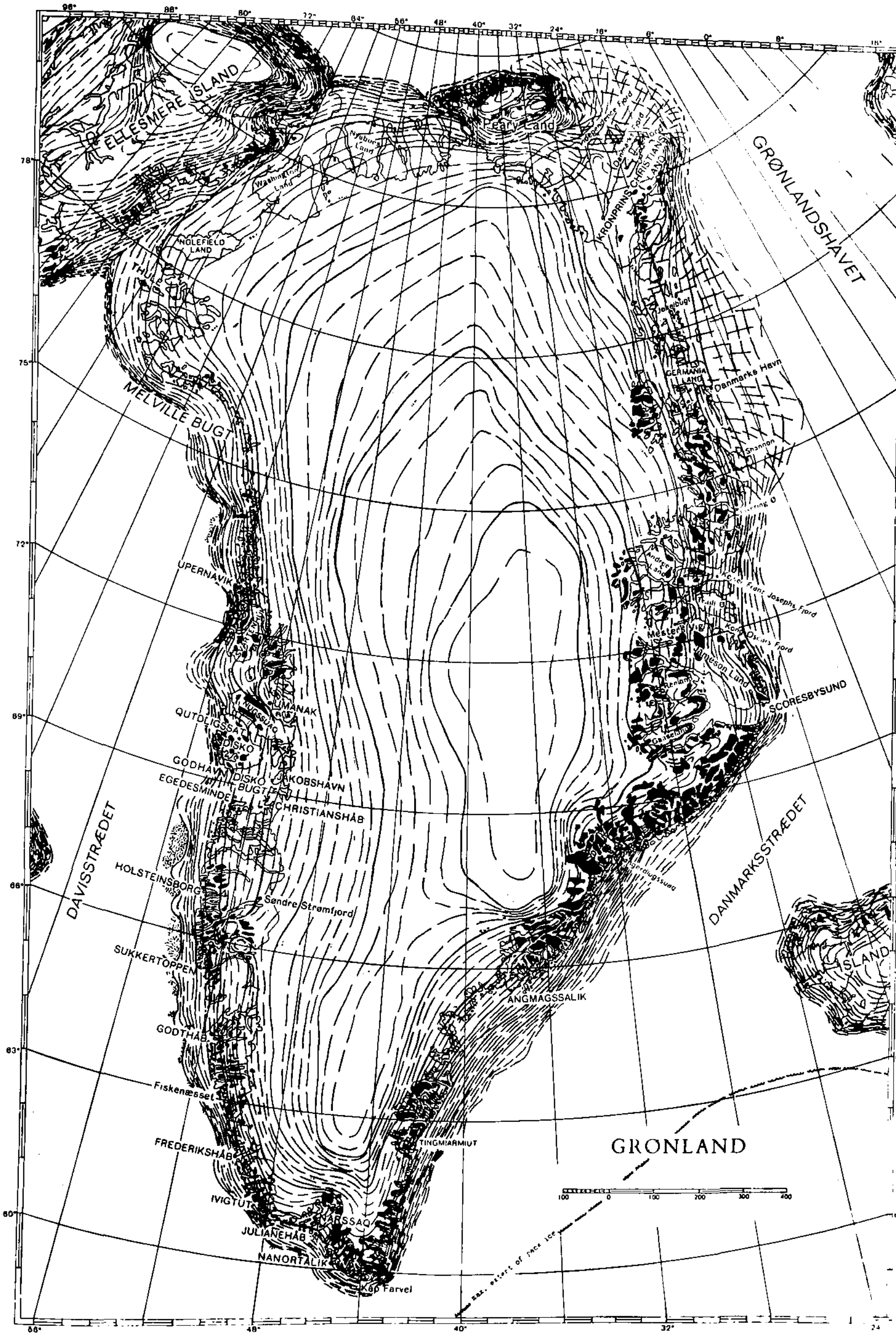


Fig. 20. Presumed extent of the Inland Ice during Wisconsin (?)time Adjoining parts of Ellesmere and Baffin glacier complexes after Craig and Fyles (1960) and Smith (1961). Iceland after Kaiser (1969). Limits of polar pack ice at annual maximum according to Flint (1947) and Lamb (1964).

Off northeast Greenland the Inland Ice may have formed great ice shelves as it did in Peary Land. Bathymetric charts show large parts of this area to have depths less than 100 m, but the mapping is not sufficient to trace details of the submarine landscape, and the ice margin ended on dry land.

Off southeast Greenland between Scoresby Sund and Kap Farvel, both the thresholds barring fjord troughs from the continental slope and the morphology of the banks may be wholly or partly explained as moraines. If this explanation is true, the Inland Ice margin between Scoresby Sund (Fig. 21) and Kap Farvel extended only slightly beyond the coast during the ice ages.

Glaciation limits

A determination of the glaciation limits (cf. p. 15) during the glacial times based on the height of cirque floors has not yet been made in detail for Greenland. Even in South Greenland cirques are found at the outer coast with floors near present sea level. Assuming that these cirques (cf. p. 53) are an expression of the elevation of the glaciation limit during the Wisconsin, the glaciation limit must have been at least 600 to 700 m below its present elevation.

While the lowering of the glaciation limit in temperate climates during the Wisconsin/Würm glaciation was 1,000 to 1,200 m (Kaiser, 1969), depression of the glaciation limit is thought to have been only half as great in arctic areas (Klute, 1928; Frenzel, 1967), which agrees with the above observations of a 600 to 700 m depression in Greenland. This implies that the firn line of the Inland Ice was only a few hundred meters above sea level even in the southernmost inland parts of Greenland. Thus only a small part of the ice cover would have been in the ablation area, and most ablation would have been by calving, even if the accumulation decreased to one-half of that at the present time.

Nunataks

The presumed ice age extent of the Inland Ice shown in Fig. 20 is based on data referred to in the foregoing pages. Local glaciations must have occurred in the highlands of East Greenland and in minor parts of West Greenland that protruded above the Inland Ice surface as well as in Peary Land.

The alpine peaks of many of the highland areas and their coastal lee slopes probably existed as small ice-free areas. An exact delineation of these areas cannot be given on the basis of the current data and their possible locations are therefore shown in Fig. 20 as "potential nunatak areas".

Fig. 21. Entrance of Scoresby Sund, East Greenland, seen from the north. In the foreground Jameson Land with widespread moraine at the coast reworked by former higher sea levels. Geodetic Institute's route 682 Q-S no. 45 (16.07.1961). Copyright Geodetic Institute.

The existence of nunataks as refugia for plants has been advocated by Bøcher (1949, 1956, 1959) in order to explain present plant distribution. This idea was opposed by Iversen (1953), who explained present plant distribution in other ways such as the spreadings of seed by ocean currents (drift ice) or birds.

In any case, the size of ice age nunatak areas must have been very restricted by strongly developed local glaciations.

The documentation of a partially ice-free outer East Greenland during at least the last part of Würm (Wisconsin) (Fig. 21) time (Funder and Hjort, 1973) may give support to the hypothesis of nunatak or semi-nunatak refugia in East Greenland. However, it must still be emphasized that an older glaciation also has covered the outer coastal stretch and, as shown by Bøcher (1956), one of the endemic plant species (*Potentilla rubella*) occurs in the inner central parts of the coastal stretch, i.e. in areas which without doubt have been covered by ice up to Holocene time.

HOLOCENE MARINE FEATURES AND DEPOSITS

The isostatic uplift of the coastland that was a consequence of the retreat of the Wisconsin ice cover is recorded in numerous marine features and deposits in much of ice-free Greenland. Comprehensive mapping and dating of former beaches are required to link together Holocene events such as phases of deglaciation and development of climate throughout the relatively large ice-free areas, but such investigations are still in a very early stage for making general conclusions.

Erosional features

Evidence of marine erosion occurs mostly in the form of coastal terraces developed on unconsolidated sediments, and rarely as terraces in bedrock. Terraces in bedrock on the west and east coasts of Greenland are described by Eberlin and Knutsen (1889), Noe-Nygaard (1932), Poser (1932) and Belknap (1941). Eberlin and Knutsen supposed that formation of an ice foot along the coasts had a strong erosional effect in East Greenland. From North Greenland, Koch (1928a) gave detailed descriptions of this phenomenon, but did not discuss its erosional effect.

Marine terraces are best observed in coarse material such as gravel or boulders, which probably have a greater resistance to solifluction than finer sediments. Thus, well-developed terraces often occur on raised deltas where a gradual transition between fluvial and deltaic terraces is common.

Depositional features

Beach ridges are especially well developed in morainic material in bays near the outer coasts, where they have been formed mostly by redeposition. In southern West and East Greenland, determination of marine levels is almost exclusively based on the height of raised beach ridges. In North Greenland also, the uppermost marine levels are characterized by unfossiliferous marine deposits or beach ridges. The lack of subfossils to be found in the uppermost deposits, as in Norway, can best be explained by low salinity or close proximity to the ice margin during deglaciation.

While beach ridges and terraces are often distinct, they have occasionally been confused with moraines or other glacial features even though the possibility of this error was mentioned by Jessen (1896). Such confusion is particularly troublesome in recognizing the uppermost marine levels.

Shell banks have been found in Greenland in situations similar to those described from Scandinavia, and as in west Sweden, such shell banks can contain mixed faunas from different periods. Shell banks are most common along the west coast between Godthåb and Egedesminde districts, and lower parts of the towns of Holsteinsborg and Sukkertoppen are built on extensive shell banks. Some localities were described from the west coast by Laursen (1950) and from the east coast by Noe-Nygaard (1932).

Clay-silt terraces with marine shells are possibly the most frequent shell-bearing deposits in West Greenland, occurring mostly in the interior of the fjords. Their lithology and mode of formation in West Greenland has been described by Laursen (1950), and in East Greenland by Sugden and John (1965). The greatest areas with such deposits seem to be Egedesminde district of West Greenland, where the Naternaq plain ("ŋaternaq Greenlandic", "like a floor") covers an area of about 250 km^2 (see Fig. 22). Parts of this plain have been described by Rink (1853), Hammer (1889) and Harder _et al_. (1949).

Areas of a similar nature as the Naternaq plain occur as far south as Godthåb district, where they have been described by Kornerup (1879, 1881) and Jensen (1889). In some places the clay deposits seem to extend under the Inland Ice (Jahn, 1938). This, together with the occurrence of concretions along the Inland Ice margin in West Greenland (cf. p.), raises a debate as to whether the terrace materials are of interglacial age, Holocene age, or interglacial age redeposited during the Holocene. It seems from the clear sequence of the molluscan fauna in many of these deposits that they must be of Holocene age, although the above mentioned possibilities still should be considered.

In North Greenland numerous deposits similar to those in West Greenland have been described from the Thule area, Inglefield Land, Hall Land and Peary Land (Davies, 1959, 1961; Davies _et al_., 1959; Nichols, 1969). In East Greenland, similar localities have been reported from the northeast and central coasts (A.S. Jensen, 1917; Wyllie, 1957; Flint, 1948; Washburn and Stuiver, 1962; Lasca, 1969; Sugden and John, 1965; Funder, 1970).

Shore-line displacements

While marine limits, in accordance with the gradual deglaciation of the land, must be metachronous and only furnish evidence of the magnitude of former depression, lower isobases supply evidence about the nature of uplift related to glaciation. Much has been written about this relationship and for general theory the reader is referred to Andrews (1966, 1968, 1969), Andersen (1965) Lundquist (1965), Donner (1969) and Ten Brink (1971).

Fig. 22. Naternaq plain, West Greenland, seen from the west. In the background Nordenskiölds Gletscher and the Inland Ice. Geodetic Institute's route 510 E-Ø no. 35 (19.08.1948). Copyright Geodetic Institute, Copenhagen.

The uppermost marine level

The description of the uppermost marine level (marine limit) must be treated with reservation because of the nature of observations and the possible confusion between non-marine and marine features. While the criteria for identifying former sea levels are primarily the presence of undisturbed shell-bearing beds and well-developed beach ridges, these can be applied in only a few cases for the uppermost marine level. Therefore, the criteria of the lower limits of undisturbed moraine and perched boulders (Sim, 1960) have also been utilized. Furthermore, these criteria must be applied to a specific locality and cannot be used to give a general statement for large areas.

With application of the criteria given above, many of the highest values for marine limit such as those given by Bessels (1879), Freuchen (1915), Koch (1917) Bøggild (1928), Koch (1928a, b), Backlund (1931b), Poser (1932) and Laursen (1950), have been deleted, and only the following altitudes can be listed (Table 6).

Table 6. Upper marine limit in Greenland

Region	District or area	Altitude, m	Source of information
West Greenland			
	Julianehåb	50-60	Jessen (1896), Bøgvad (1940), Weidick (1963)
	Frederikshåb	94	Jessen (1896), Kelly (1966)
	Godthåb	100-110	Kornerup (1879), Iversen (1953)
	Holsteinsborg	110-140	Weidick (1968a, b), Weidick and Ten Brink (1970)
	Egedesminde	108-150	Pjetursson (1898), Kelly (1969)
	Disko Bugt	130-150	Steenstrup (1883b), Laursen (1950) Donner (1973)
	Umanak	130-200	Steenstrup (1883b), Laursen (1950)
North Greenland			
	Thule	50	Davies *et al.* (1963)
	Carey Øer	90	Bendix-Almgreen *et al.* (1967)
	Inglefield Land	below 100	Nichols (1969)
	Peary Land	130	Troelsen (1952a), Davies (1963)

Table 6 (continued)

East Greenland			
	Fjord region	216-100	Bretz (1935), Noe-Nygaard (1932)
	Mesters Vig	120	Washburn (1965), Lasca (1969)
	Scoresby Sund	134	Sugden and John (1965), Funder (1972)
	Angmagssalik-Kap Farvel	c.75	Vogt (1933)

Despite their spread, the data show a uniform trend; i.e. except for South Greenland and possibly North-West Greenland, most maximum values are 110-150 m. The lowest marine limits occur close to the outer coast, where they may be below 100 m; and in wide ice-free areas, similarly low limits again occur close to the present Inland Ice margin. The only unusually high marine limit is from the Fjord region, where Noe-Nygaard in Vega Sund found "the highest unmistakable marine limit" at 216 m altitude and the highest fossiliferous strata at 138 m. In Umanak district, West Greenland, Laursen described shell-carrying clay beds at 190 and 200 m, but subsequent dating of neighboring "interglacial" shell beds may make it doubtful that these deposits are of Holocene age (cf. p. 50).

The general trend of the highest marine limit suggests a Wisconsin ice cover that left only tiny areas ice-free.

Marine levels below the uppermost

A great amount of information on raised shorelines in Greenland was collected by Vogt (1933), Bretz (1935) and Laursen (1944, 1950, 1954) in order to discern specific levels. While the first two authors essentially based their correlations on morphology, Laursen also included a faunal stratigraphy.

In West Greenland a reportedly pronounced terrace with an average altitude of about 10 m is presumed to have been formed at the end of the climatic optimum (Laursen, 1950). With regard to higher marine levels, Laursen (1944) made the following correlations between West Greenland features and those reported from North Greenland by Koch (1928a, b) (Table 7).

Table 7. Correlation of marine levels in West and North Greenland (according to Laursen, 1944)

Disko Bugt	Umanak district	Kap York area	North Greenland
	85-101 m		105 m (80-110 m)
	21- 45 m		65 m (45- 57 m) (25- 37 m)
10 m	10 m		

The figures in parentheses alongside those of Laursen and Koch refer to subsequent measurements by others of presumably the same levels. A major level at about 40 m occurs around Kap York according to Davies et al. (1963), and higher levels of about 110 m and 80-100 m occur respectively in Hall Land and Jørgen Brønlund Fjord according to Davies (1963).

It is not quite clear from comparison between the diagrams of Troelsen and Davies (Troelsen, 1952a; Davies, 1963), whether Davies' 45-57 m level in Peary Land is identical with the 65-m level of Koch and Troelsen, though it would appear to be so because both levels are described in connection with a glacier advance through the fjord. Davies found the major zone of marine levels north of Jørgen Brønlund Fjord to occur at 45-57 m, while farther south the same zone had an altitude of 25-37 m. However, secondary levels on his diagram can also be discerned around 65 m and 80-100 m in the Independence Fjord and Jørgen Brønlund Fjord area and while the 65-m terrace according to Troelsen was formed in connection with the above-mentioned glacier advance through Jørgen Brønlund Fjord, Davies gives generally for Peary Land a glacial advance just prior to the formation of terraces 24 to 50 m above present sea level.

For North-East Greenland, Bretz (1935) compiled all available information about marine levels in various areas. Despite the great number of observations, he concluded (ibid. p. 219): "A glance at the table shows the hopelessness of recognizing any definite level in more than half of the studied localities and the large number of cases where only one, two, or three localities contain a record".

A treatment by Vogt (1933) especially concerned a pronounced marine level in South-East and West Greenland regarded by Vogt as the Greenland equivalent of the Fennoscandian Tapes line, the maximum age of which according to Andersen (1965) is 7860 years B.P. In Greenland, the average altitude of this level was given as 25 m at the outer coasts and 35-40 near the Inland Ice margin.

Age determination of former sea levels in Greenland

Former shorelines have been dated by radiocarbon analysis of shells and driftwood, and by the determination of the times when lakes were isolated from the sea through investigation of the lake sediments. Most radiocarbon dates have been derived from three areas of Greenland:

(1) West Greenland (Fredskild, 1967; Weidick, 1968a, 1972; Kelly in Tauber, 1968; Ten Brink, 1971, 1974).

(2) Northeast and eastern North Greenland (Davies in Rubin and Alexander, 1960; Davies in Ives et al., 1964; Davies in Trautman and Willis, 1966; Davies, 1963; Knuth in Tauber, 1960a, b, 1961, 1964, 1966a, b, 1968; Knuth, 1963, 1964; Fredskild, 1969a).

(3) Mesters Vig area, East Greenland (Washburn and Stuiver, 1962; Lasca, 1969) (cf. Fig. 25).

In determining the age of former sea levels, errors are involved in both age determination (Tauber, 1958) and the source of carbon in the sample (Tauber, 1966b; Dyck and Fyles, 1964). Further errors may be involved in the relation of sea level to the dated material (Andrews, 1966). Dating errors may be somewhat controlled by geometric relations between the samples: the data from one locality have to fit, and emergence curve and levels of the same age at different localities inside an area have to be linked together by an isobase. Another control is provided by dating former sea levels within the same area by the different means given above (drift wood, shells, isolation of lakes).

With reservations due to the errors given above, emergence curves have been published by Weidick (1972) and Ten Brink (1974) for West Greenland and for eastern North Greenland and have been compared to those for Mesters Vig and Skeldal given by Washburn and Stuiver (1962), Washburn (1965) and Lasca (1969). They are here given in Fig. 23. They seem to indicate rather clear trends, namely that uplift occurred relatively early in the southernmost and outermost coastal areas where most of it was achieved by 4000-5000 years B.P. The amount of uplift in southernmost West and East Greenland was considerably less than in other parts of Greenland, with the possible exception of the most northern coastal parts such as Kaffeklubben Ø.

The relatively late uplift of North Greenland is especially pronounced for the inner parts of the coastland (Jørgen Brønlund Fjord, Danmark Fjord), and also seems to be indicated by the two published dates from Polaris Promontory, Hall Land (Rubin and Alexander, 1960) and Dallas Bugt, Inglefield Land (Nichols, 1969). North Greenland uplift continued until at least 2,000 years ago as demonstrated by C^{14}-dated relations of the Independence cultures (Knuth, 1963, 1964, 1967).

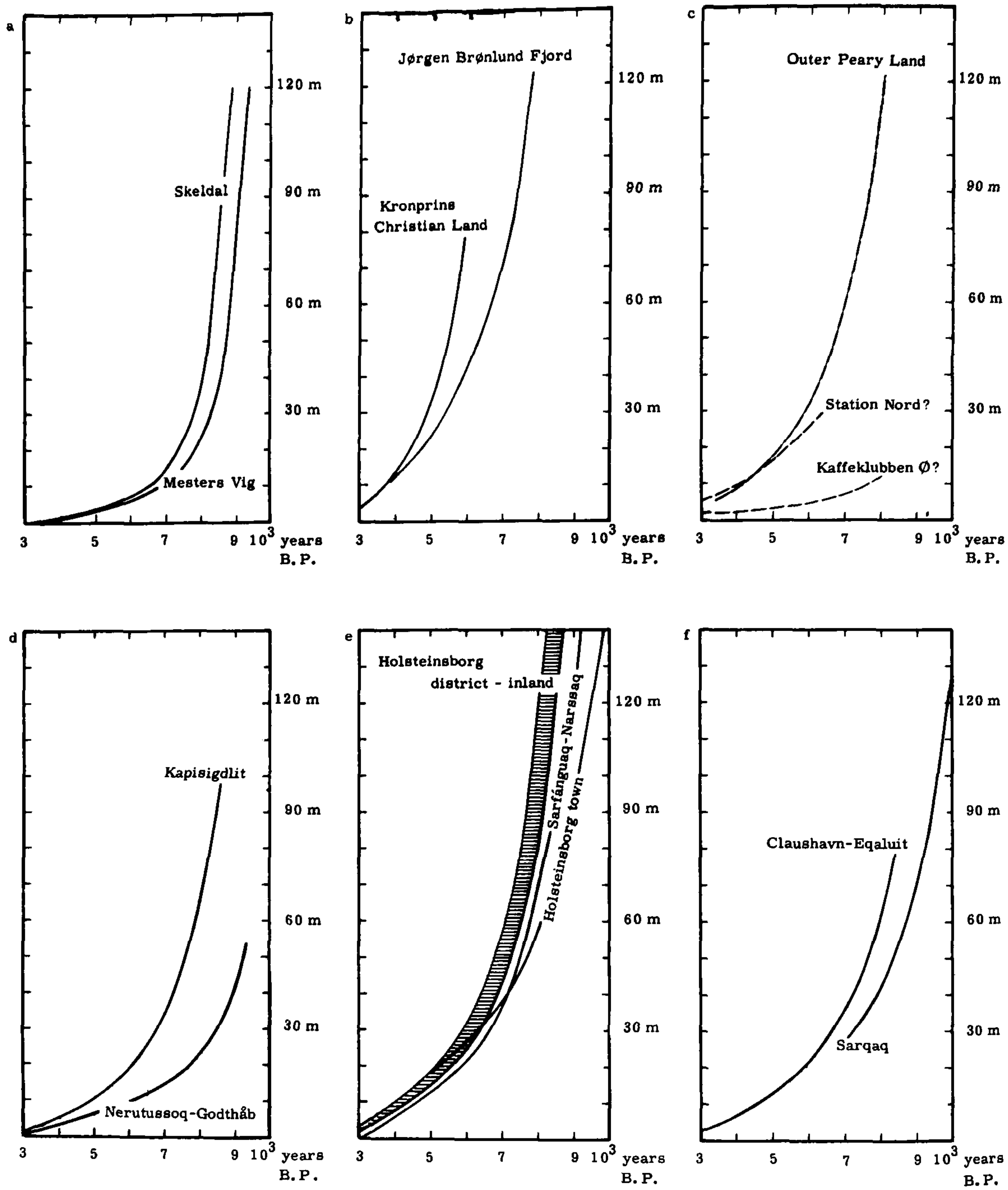


Fig. 23. Uplift curves expressed by uncorrected radiocarbon dates versus field altitudes. In West Greenland according to data of Kelly (in Tauber, 1968) and Weidick (1972). In East Greenland (Skeldal and Mesters Vig) according to data of Lasca (1969) and Washburn (1965). The curves of North Greenland (Jørgen Brønlund Fjord and Kaffeklubben Ø) present best fit of data, compiled from Davies (in Rubin and Alexander, 1960; Ives et al., 1964; Trautman and Willis, 1966), Knuth, Fredskild and Kirkeby (in Tauber, 1960a, b, 1964, 1966a, b,). As for East Greenland, the North Greenland curves are not corrected for any apparent ages, nor is the eventual break of the uplift curve of Jørgen Brønlund Fjord caused by readvance through the fjord of the Inland Ice taken into consideration. After Weidick (1972).

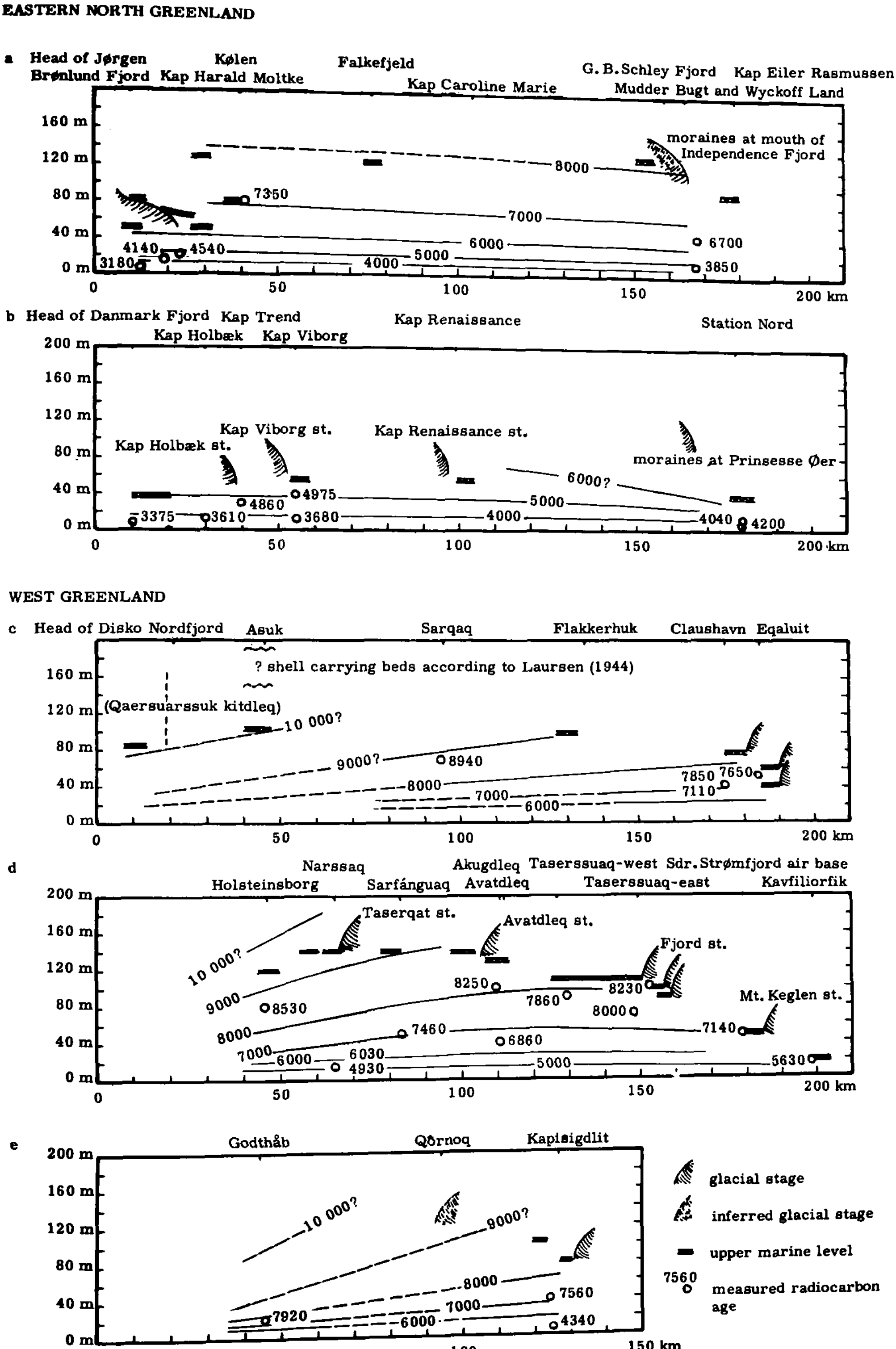


Fig. 24. Shoreline diagrams for sections of eastern North Greenland and from West Greenland. The relation of the shorelines to former stages in deglaciation is shown schematically. From Weidick (1972).

Fig. 25. Mesters Vig, East Greenland, seen from the northeast. Geodetic Institute's route 652 A-V no. 12838 (17.8.1950). Copyright Geodetic Institute.

The uplift curve of Jørgen Brønlund Fjord given in Fig. 23 differs from that given by Fredskild (1969a) for higher altitudes because of Fredskild's correction of 1,900 years based on the C^{14} age of modern lake sediments, while the presented uplift curves of Jørgen Brønlund Fjord and Danmark Fjord are based on a correction of shell samples by 1,200 years deduced by comparison with dated samples of other kinds (Weidick, 1972).

Tilt of the marine levels indicated by the current data is shown in Fig. 24. As in other glaciated regions, isobases rise toward the center of glaciation. Because Greenland is still partly covered with ice the trend of tilt close to the present Inland Ice margin is interesting, but not enough data are available in these regions for a detailed picture.

Archeological investigations have provided information about more recent fluctuations of sea level from observations in both West and East Greenland that medieval ruins are now partly or completely under water (Bøgvad, 1940). Sinking of the land apparently started in A.D. 1600-1700 and has continued until the present time (Egedal, 1947; Nielsen, 1952; Saxov, 1958, 1961). Saxov gave the rate of sinking as 6-14 mm/year between 1897 and 1958 for the Disko Bugt area, West Greenland, and for Angmagssalik in South-East Greenland he found the value to be between 2.7 and 3.4 mm/year. Furthermore, Saxov stated that from c. 1940 the sinking of the land has been replaced by uplift. Thus, the period of sinking seems in part to coincide with the most recent period of glacier readvance (see also p. 93).

A "Vega transgression" is stated in East Greenland by Hjort (1973) and is correlated to the Norwegian "Taped III" transgression (c. 5500 B.P.).

Holocene marine faunal development

Investigations of the Holocene marine fauna with special reference to the molluscs were initiated by Jensen and Harder (1910) and continued by Laursen (1950). The investigations revealed development from a high arctic fauna (Portlandia arctica, Balanus hameri) to an arctic fauna (Mytilus edulis, Pecten islandicus), followed by a boreal fauna (Zirphea crispata, Cyprina islandica) and finally the present arctic fauna. This sequence indicated correspondence between the Holocene climatic optimum and the boreal fauna. The early high arctic fauna showed a sequence of high arctic-arctic-high arctic, which Laursen correlated with the sequence Older Dryas-Allerød-Younger Dryas. On the basis of the faunas Laursen estimated post-July temperatures (Laursen, 1953). However, this correlation was done before the knowledge of pre-Hypsithermal cold periods younger than younger Dryas, and the sequence may therefore be younger.

In West Greenland and East Greenland, the appearance of *Mytilus edulis* has been taken as evidence of the beginning of the Holocene climatic optimum (Noe-Nygaard, 1932; Laursen, 1950; Hjort and Funder, 1974).

HOLOCENE GLACIAL DEPOSITS

The influence of topography and climate on the retreating ice resulted in formation of numerous Holocene moraines, especially at the mouths of tributary fjords. While their presence here is often reported locally, mapping of the margin deposits over large areas is rare. Therefore correlation of the deposits reported from widely separated areas is largely unfeasible.

Lithology and extent of the deposits

Ground moraine must be considered the most widespread Holocene deposit in Greenland. However, it is often only a thin layer less than a few meters thick, even in valleys and hollows. In lithology, the moraines vary between clayey, silty, sandy and gravelly till, the last one being the most common especially at higher elevations. In the lowlands near the Inland Ice margin in West Greenland, clayey till seems to be common, even up to altitudes of 500 m. The primary source for this kind of till may be eroded marine or freshwater clay deposits.

The boulders in the clayey and silty tills are often relatively small, mostly subangular in shape. With reference to this kind of till, it must be mentioned that Washburn and Stuiver (1962) have described a till-like fjord sediment, which was deposited by icebergs. Such "drift" can be expected at all levels up to the marine limit in Greenland.

In sandy or gravelly till the boulders as well as the coarser matrix are often well rounded. This indicates some fluvial as well as glacial transport.

Extensive areas of ground moraine seem to be restricted to inland areas of Holsteinsborg and Egedesminde districts in West Greenland, Inglefield Land, Hall Land, parts of Kronprins Christian Land and Peary Land in North Greenland, and some areas of Germania Land, Dronning Louise Land and Jameson Land in East Greenland. This distribution is shown in the Quaternary Map of Greenland.

Transitions from moraine to marine, glaciofluvial, and glaciolimnic sediments occur frequently, and widely occurring kame terraces consist mostly of coarse sediments, which frequently grade laterally into moraine or glacial lake deposits. This whole complex of phenomena are best considered as ice margin features.

Varved clay, also an ice margin feature deposit, seems to be relatively rare which may be due to the paucity of investigated areas.

Varved clay has been reported from Julianehåb district (Weidick, 1963) and Sukkertoppen district (Etienne, 1940; Sugden and Mott, 1940) in West Greenland, from the Thule area in North Greenland (White, 1956), and from Dronning Louise Land (Lister and Wyllie, 1957) and Ile de France (Bronner, 1948) in East Greenland. Furthermore, varve clay is reported deposited at the surface of the Inland Ice margin near Thule, North Greenland (Swinzow, 1962).

Morphology

The ground moraine often thickens locally and forms terminal or marginal moraines. Where the moraines are continuous over wide areas, they must indicate major halts or readvances during Holocene deglaciation (Fig. 27). Generally these marginal deposits are developed as moraine ridges at lower altitudes and as boulder ridges or attenuated moraines on slopes at higher altitudes.

Spillways and kame terraces often occur along the former drift borders and have been described from West Greenland by Milthers (1948). A large kame terrace was described from Peary Land by Troelsen (1952a) and Kirkeby (*in* Knuth, 1964), and Bretz (1935) and Flint (1948) described similar features in East Greenland.

Concentric beach ridges often encircle the sites of former and present ice-dammed lakes. The sediments on which the beach ridges are formed show a wide variation between typical moraine and freshwater deposits, which can often be studied even at present ice-dammed lakes (Weidick, 1963; Higgins, 1970).

Eskers were first described from West Greenland in the Holsteinsborg district (Kornerup, 1881; Milthers, 1948). A review of aerial photographs has revealed widespread occurrences of this feature, the older of which are mostly in middle West Greenland, while younger features occur in areas deglaciated during this century at many places along the coast (Weidick, 1968a; Henderson, personal communication).

Dating of the ice margin deposits

An attempt to date several individual stages has been made in West Greenland by referring local glaciations to different glaciation limits (Weidick, 1963, 1968a) and by correlating former sea levels with the formation of moraines, such as those shown in Fig. 24.

Varve countings have been tried for restricted areas in Thule, North Greenland (Goldthwait *in* White, 1956) and in North-East Greenland

(Lister and Wyllie, 1957), but have covered events of only the last 140 and 2,000 years, respectively. Restricted time-coverage also limited dating of moraines by lichenometry in West Greenland (Beschel, 1961) and South-East Greenland (Gribbon, 1964) as well as the tree-ring analyses attempted in West Greenland (Jakobsen in Weidick, 1963) and North Greenland (White, 1956). While tree-ring analyses can only be applied to events within this century, lichenometric dating may have a working range of 500-1,000 years.

In addition to the above methods, written records have been used to date recent glacier fluctuations (Koch, 1928a; Davies and Krinsley, 1962; Weidick, 1968a, 1969). Even in the most favorable circumstances historic records extend back to only the beginning of the 19th century for Greenland.

Ice margin stages in West Greenland

In the account below, short summaries are given for West, North and East Greenland, and within each region ice margin deposits are roughly classified into stages older and younger than the Holocene climatic optimum. Because most isostatic recovery occurred prior to the Holocene climatic optimum in most parts of West and East Greenland, dating of earlier stages by their relations to former sea levels is largely restricted to moraines deposited prior to (or during?) the climatic optimum ("older moraines"). Younger moraines can only be related to former sea levels in North Greenland, and elsewhere their age is estimated by one of the methods summarized above.

Older ice margin deposits

The oldest moraines described are the presumed Wisconsin nunatak moraines in the outermost western parts of the coastland (cf. p. 57). A younger moraine zone ("the nunatak zone") occurs in the inner parts of Amerdloq and Ikertoq fjords of Holsteinsborg district and comprises at least two stages (Avatdleq and Taserqat) as shown in Fig. 28 (Weidick, 1968a,b). The formation of these moraines contemporaneous with a sea level close to marine limit for the area implies an age of around 10,000 years B.P. and 8,700 years, respectively. A northern continuation of these ice-margin deposits into the southern parts of Egedesminde district was described by Kelly (1969).

A subsequently developed major zone of ice-contact deposits ("the outer zone") can be traced from 64° to 70° N, roughly parallel to the present Inland Ice margin at distances of 5-40 km from the ice margin.

The outer zone can be divided into several local stages, and recent radiocarbon datings of related sea levels furnish the following results at three sectors of the outer zone (Weidick, 1972) (Table 8).

Table 8. Approximate ages and related sea levels to stages in the recession of the West Greenland Inland Ice margin

Disko Bugt		Holsteinsborg district		Godthåb district	
m a.s.l.*	age	m a.s.l	age	m a.s.l.	age
35	6800-7200 B.P.	45-50	7100-7300 B.P.		
60	7600 B.P.	90	8100 B.P.		
70-80	8100-8300 B.P.	100	8300 B.P.	80	8300 B.P.
		110	8400 B.P.		

* approximate former sea level at formation of the ice margin deposits.

All the ages must be considered as approximate. In Holsteinsborg district the three oldest stages, similarly those in Disko Bugt and in Godthåb district, have been designated fjord stages while the youngest one in Søndre Strømfjord has been named the Mt. Keglen stage.

A southern continuation of the outer zone was reported from Frederikshåb district at 62° N (Kelly, 1966). A minimum age furnished here is 9580 B.P. for a western stage in deglaciation (Kelly in Tauber, 1968). However, the paucity of ice margin deposits in the intervening areas around 63° N led Graff-Petersen (1952) to conclude that rather continuous recession of the Inland Ice occurred throughout the area.

In the southernmost parts of West Greenland, Weidick (1963) has distinguished two stages supposedly older than the climatic optimum: the Niaqornakasik and the Tunugdliarfik stages. The Niaqornakasik stage is related to a relative sea level of approximately 30 m and the Tunugdliarfik stage is related to a sea level 10-15 m above the present. However, lack of knowledge about age of former sea levels in this area makes correlation with stages farther north highly tentative.

Younger ice margin deposits

During the climatic optimum the Inland Ice and local glaciers must have been less extensive than at present. Evidence of this is the occurrence of marine Holocene deposits under the margin of the Inland Ice in Egedesminde and Holsteinsborg districts and under local glaciers in the Sukkertoppen district. In Disko Bugt shear moraines at

Pakitsoq contained branches of <u>Betula nana</u> carried up from the substratum of the Inland Ice and the shell and concretion carrying moraines found at the Inland Ice margin in much of West Greenland (Tarr, 1897b; Gripp, 1932; Weidick, 1971b) may also indicate formerly less extensive glaciers (cf. p. 50).

Ice margin deposits formed at present sea level occur frequently inside the outer zone along the Inland Ice margin and around local glacier lobes. The soil and vegetation developed on the deposits indicate that they were formed in the early parts of the "little ice age" and before "historic time". The term "little ice age" is used here in the sense of Matthes (1942); i.e., covering approximately the last 4,000 years and "historic time" for Greenland indicates the period of <u>c.</u> 1600-1920 A.D.

At Qajá in Jakobshavns Isfjord, archeological sites of the Sarqaq culture (age around 3500 B.P., cf. p. 99) are developed inside younger ice-margin deposits. Thus for at least that locality a minimum age is known for the moraines which is coincident with the early part of the little ice age.

With reference to glacier fluctuations in historic time, lichenometric and literary evidence indicates general similarity between events in Greenland and those reported from Europe (Beschel, 1961; Weidick, 1968a). For certain glaciers in the Holsteinsborg district Beschel expanded this parallelism to the whole little ice age.

Ice margin stages in North Greenland

A variety of investigations in North Greenland have furnished rather controversial data, so that any synthesis of the data is questionable.

Older ice margin deposits

Moraines formed by a readvance through the fjords were reported from Hall Land by Koch (1928a). Shells in marine deposits distal to the moraines, pre-dating the advance, had an age of 6100 years B.P. (W-816) and shells from a terrace cut into the moraines (W-815) had an age of 3780 years B.P. which supplied a minimum date for the advance (Rubin and Alexander, 1960; Davies, 1963). The deposits are shown here in Fig. 26.

In Peary Land and Kronprins Christian Land a great number of ice-contact features have been mapped by Davies and Krinsley (Davies, 1963; Davies and Krinsley, 1962; Krinsley, 1961) (Fig. 29). Shell material from a beach ridge 11 m a.s.l. at Kaffeklubben Ø near Kap Morris Jesup

Fig. 26. Hall Land and Newmann Bugt, North Greenland, seen from the north. Moraine system truncated by shorelines. Geodetic Institute's route 546 B-S, no. 11702 (22.07.1953). Copyright Geodetic Institute.

Fig. 27. Karrats Fjord in Umanak district, West Greenland, seen from the south. On the northern shore of the fjord can be seen moraines, the continuation of which across the fjord is seen as an iceberg bank. The age of the moraines is not known. Geodetic Institute.

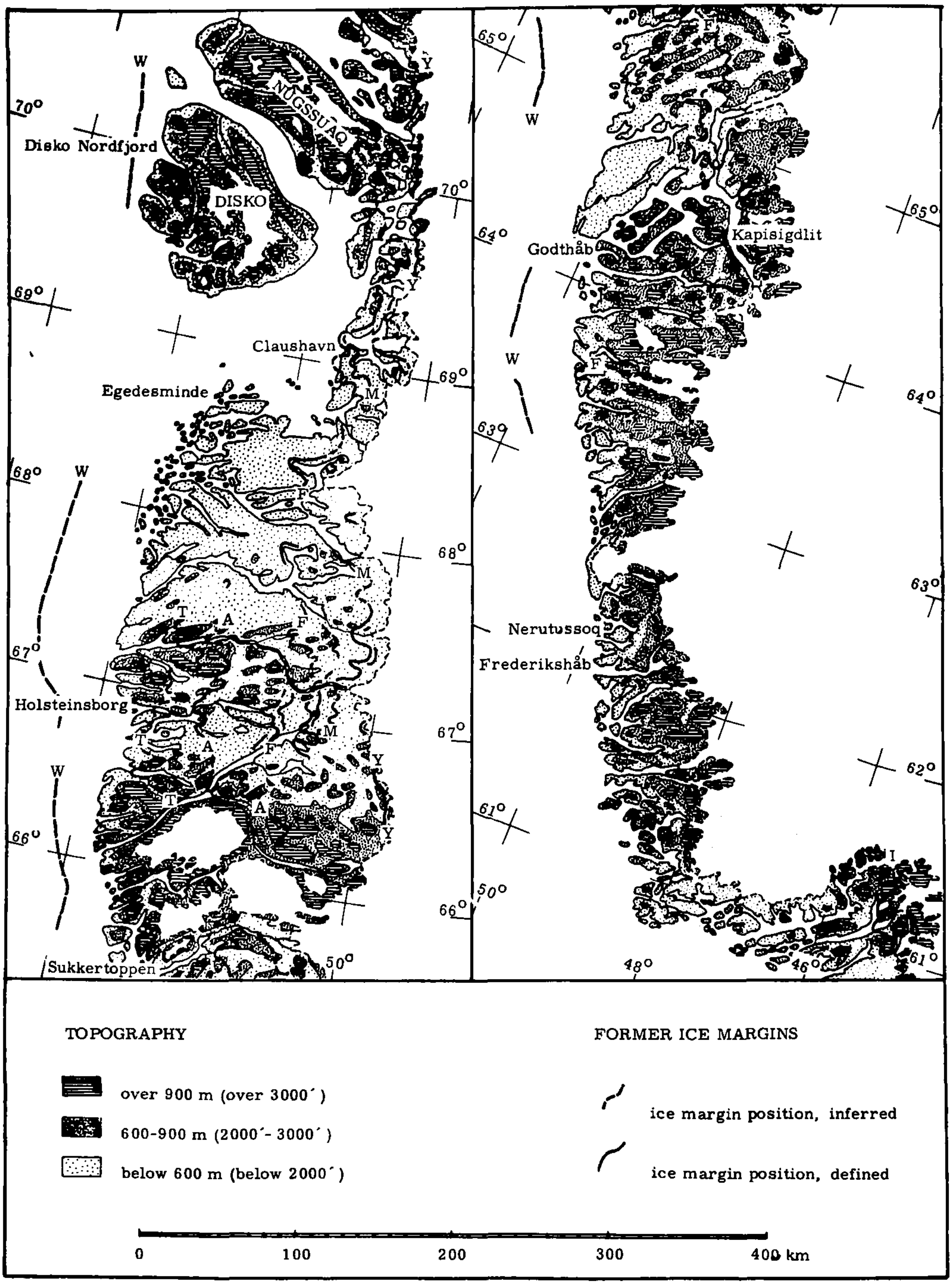


Fig. 28. West Greenland and South Greenland. Altitudinal conditions and stages in the deglaciation of the coastal area (simplified). Stages indicated are: W = Wisconsin (?), T = Taserqat, A = Avatdleq, M = Mt. Keglen, Y = Undifferentiated stages younger than Mt. Keglen, F = Fjord and I = Tunugdliarfik stage. From Weidick (1972).

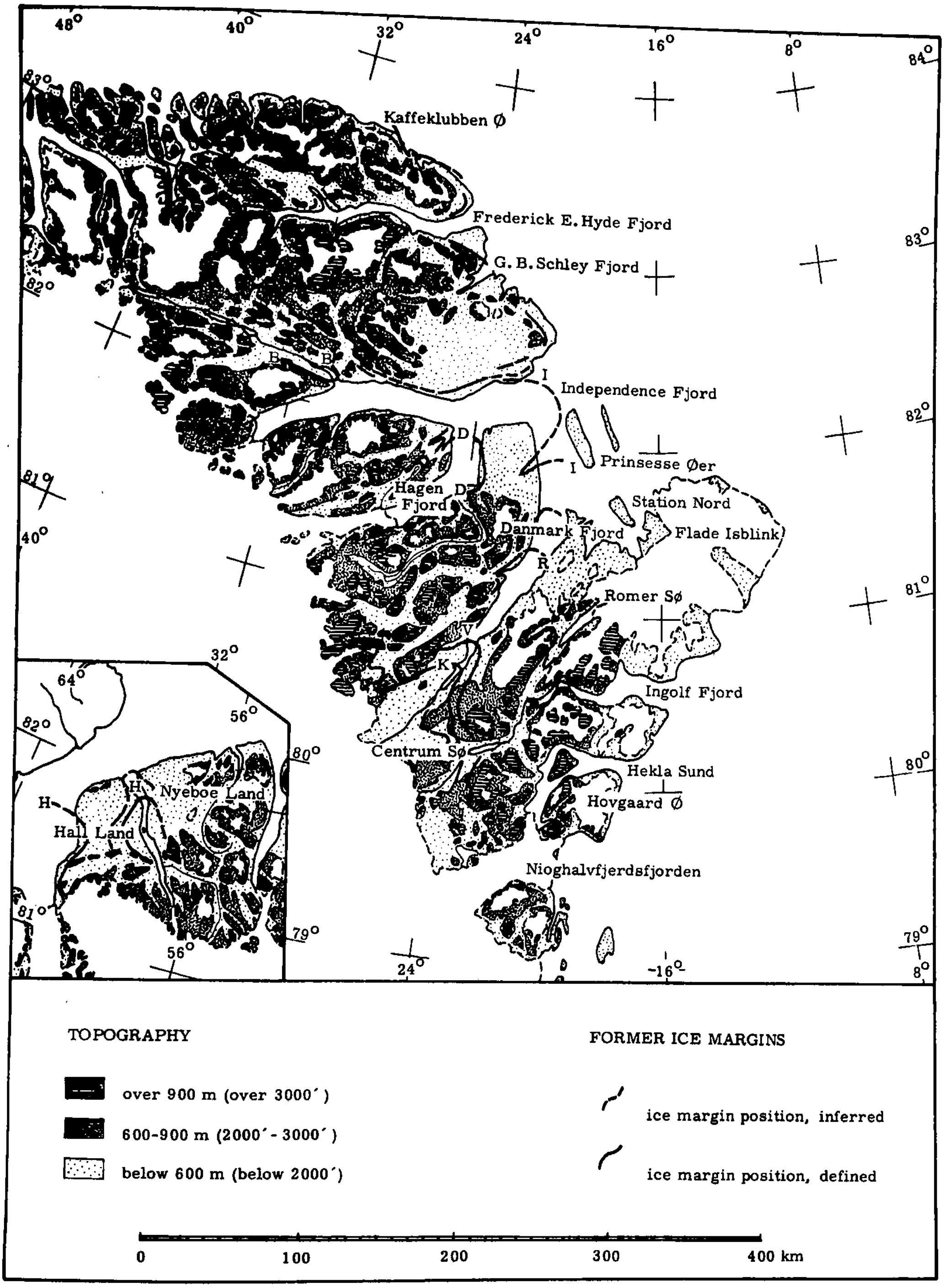


Fig. 29. North Greenland. Altitudinal conditions and former stages in the deglaciation of the coastal area (simplified). Stages indicated are: I = Mouth of Independence Fjord, B = Jørgen Brønlund Fjord, V = Kap Viborg, K = Kap Holbæk, R = Kap Renaissance, D = Deichmanns Øer, H = Hall Land. From Weidick (1972), and based on information of Davies (1963) and Krinsley (1961).

had a radiocarbon age of 7730 years B.P. (Davies and Krinsley, 1962). The uppermost beach ridge is here situated about 23 m a.s.l. on moraines, deposited by local piedmont glaciers. Therefore, the age must be considered as a minimum date for formation of the moraines.

In Peary Land, the outermost and oldest moraines are those of Wisconsin age referred to on p.50 and 62. Possibly younger moraines at the mouth of the fjords indicate readvances or halts in the ice recession correlative to those around 8000 years B.P. in West Greenland. After receding beyond its present extent, glacier lobes readvanced through the major fjords. According to Davies (1963), this readvance was about 30-60 km down fjord, and disrupted the terrace-forming stage just prior to the formation of terraces in the zone 24 to 50 m above present sea level. Since that time retreat occurred until 500 years ago, after which only minor readjustments in the position of the glacier fronts have occurred."

A readvance in Jørgen Brønlund Fjord has also been described (Troelsen, 1952a; Kirkeby _in_ Knuth, 1963, 1964). The readvance in Jørgen Brønlund Fjord seems to have been contemporaneous with a marine level of 60-70 m, and after the readvance marine terrace were formed continuously down to present sea level. According to Kirkeby, undisturbed shells of _Mya truncata in situ_ from the surface of a terrace at 40 m a.s.l. gave an age of 7740 years B.P. (K-965, _ibid_. 1964). This age must very nearly date the former sea level, as the oldest freshwater deposits on the same terrace gave an age of 6850 years B.P. (Fredskild _in_ Knuth, 1964; Fredskild, 1969a). However, the freshwater deposits were calcareous and Fredskild estimated the date to include an apparent age of 1,900 years. In the same way, most shell dates of Peary Land and Kronprins Christian Land show an apparent age of around 1,200 years compared with those of driftwood and charcoal related to the same sea level (Weidick, 1972). It is believed that the shells were dated too early due to the surrounding rocks. A dating of former morainal stages as best interpreted with relation to the current information on sea levels and datings and including a correction of an apparent age of 1,200 years on shell dates gives the following (Weidick, 1971b, 1972):

Moraines at entrance of Independence Fjord .... _c_. 8200 B.P.
Readvance of Jørgen Brønlund Fjord............. _c_. 6000 B.P.
Moraines in central and inner parts of Danmark Fjord (Kap Renaissance, Kap Viborg, Kap Holbæk and around Hall Land. . . . . 6000-5000 B.P.

Younger ice margin deposits

Ice margin deposits that by their position must be older than historic moraines have been reported from Nunatarssuaq in the Thule area (White, 1956; Davies _et al_., 1963), Brother John Gletscher in

Inglefield Land (Koch, 1928a), and as an extensive system of recessional moraines from Kronprins Christian Land (Davies and Krinsley, 1962; Krinsley, 1961). These moraines may have been formed in an early stage of the little ice age.

An advance in historic time has been described from many localities. A hint of its beginning is given by the 520 years B.P. radiocarbon date of moss from shear planes in the ice cap margin at Nunatarssuaq (W-537, Goldthwait, 1961). An extensive amount of data has been collected from North Greenland by Koch (1928a) and Davies and Krinsley (1962), which indicate that advances of many glaciers occurred up to 1920 A.D. In many cases the late advances must have been maximum for the little ice age as they eroded old, high marine terraces. Since around 1920 glacier retreat has occurred, especially in western North Greenland. However, the Inland Ice glacier lobes of Peary Land seem to have been relatively more stable.

Ice margin features in East Greenland

Older ice margin deposits

Ice margin deposits indicating an advance through the fjords around Clavering Ø (74° 30' N) were described by Gelting (1934). This Zackenberg-Finsch Island stage seems to have been formed close to a relative sea level of around 110 m, which may correlate it with the morainal stages of supposed Younger Dryas or Pre-Boreal age in Skeldal near Mesters Vig and in central Scoresby Sund (see below).

Investigations farther south around Mesters Vig have revealed that the area was open to the sea, and therefore deglaciated at least in part by 8500-9000 years B.P. (Washburn and Stuiver, 1962; Pessl, 1962; Washburn, 1965; Lasca, 1969). Prior to that time deglaciation in Skeldal was interrupted by a readvance that reached Kong Oscars Fjord.

From the northern part of Scoresby Sund in Schuchert Dal, Sugden and John (1965) described evidence of a glacial stage that interrupted glacier recession around 10,000 years B.P. and was contemporaneous with a relative sea level of 101 m. Farther south, in Hall Bredning (see Fig. 30), extensive moraine systems were observed by Bay (1896) at Danmark Ø, and they have been traced along the western shores of Renland (Funder, 1970). Bay observed that the moraines at Danmark Ø were covered by marine deposits up to 64 m. Funder has since divided the moraines into two groups related to relative sea levels of 110-120 m and 90 m, respectively. These Milne Land stages seem to be of Younger Dryas or Pre-Boreal age and are possibly to be correlated with the readvance at Schuchert Dal and through Skeldal (Funder, 1972).

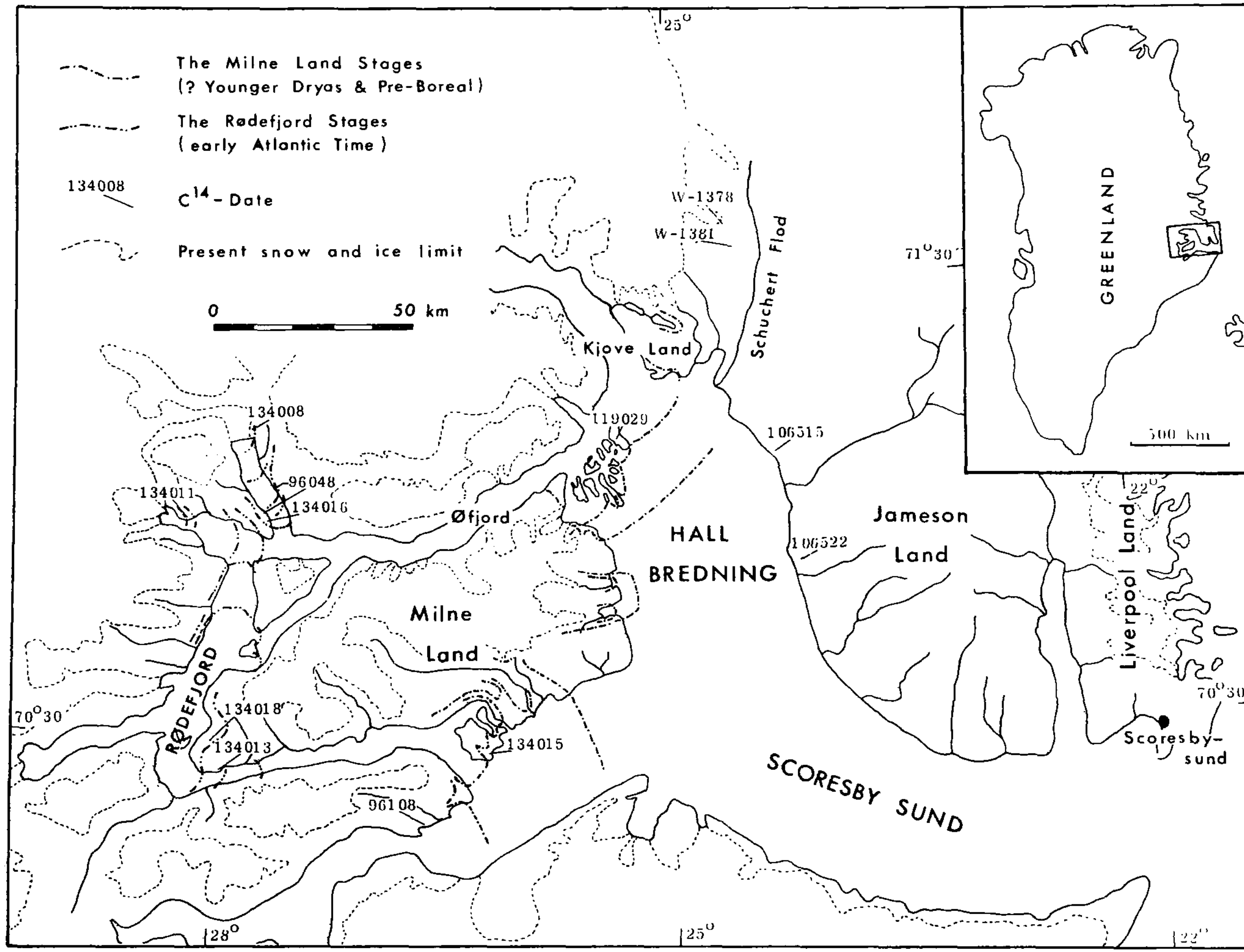


Fig. 30. Stages in the recession of the Inland Ice in Scoresby Sund. From Funder (1972).

Younger ice margin deposits

That the glaciers in East Greenland were previously less extensive seems to be indicated by shell-carrying moraines at the Inland Ice margin near Storstrømmen (A.S. Jensen, 1917). Also, in the innermost parts of Scoresby Sund marine shells were found in the moraines or Rolige Bræ (Bay, 1896), and in Schuchert Dal on the north coast of Scoresby Sund peat buried in drift from Ivar Baardsons Gletscher gave a C-14 age of 1490 years B.P. The glacier rests on marine deposits 50 m a.s.l. with shells dated to 7900 years B.P. (Schafer _in_ Levin _et al._, 1965).

Apart from Schuchert Dal there are very few observations indicating readvances in East Greenland in the early part of the little ice age. In Dronning Louise Land Lister and Wyllie (1957) concluded that nearly continuous deglaciation occurred through the last 2,000-3,000 years. Possibly some of the moraines observed in the inner parts of East Greenland fjords (Koch, 1917; Koch and Wegener, 1917; Bretz, 1935; Ahlmann, 1941; Flint, 1948) and in the inner parts of Scoresby Sund (Ryder, 1896) may therefore be of the little ice age. However a large glacier at the head of Mesters Vig is only about 8 km from an emerged delta with shells dated at 8480 B.P. (Washburn and Stuiver, 1962) and the general conclusion for investigations on the deglaciation of the inner Scoresby Sund region is that a rather continuous recession of the Inland Ice occurred between Pre-Boreal and late Atlantic time. In late Atlantic time the glaciers reached beyond their present extent and the subsequent readvance seems only to be indicated by fresh moraines, presumably of historical age (Funder, 1972).

An advance in historic time was recorded in East Greenland by several glaciers and lobes from the Inland Ice margin (Thorarinsson, 1952; Ahlmann, 1948; Sharp, 1956). The information available reveals fluctuations of the glaciers in historic time comparable to those in West Greenland (Weidick, 1968a).

Concluding remarks

In Fig. 31 the data of Greenland glacial stages in the recession of the Inland Ice have been compared to the Camp Century climatic record. It is seen that the glacial stages are comparable only in part to cold spells of the climatic record, which is true even for corrected radiocarbon ages.

The given ages reveal the possibility that the oldest known stages in West Greenland (Taserqat) and in East Greenland (around the Mesters Vig area and the Milne Land stages in Scoresby Sund central parts) are

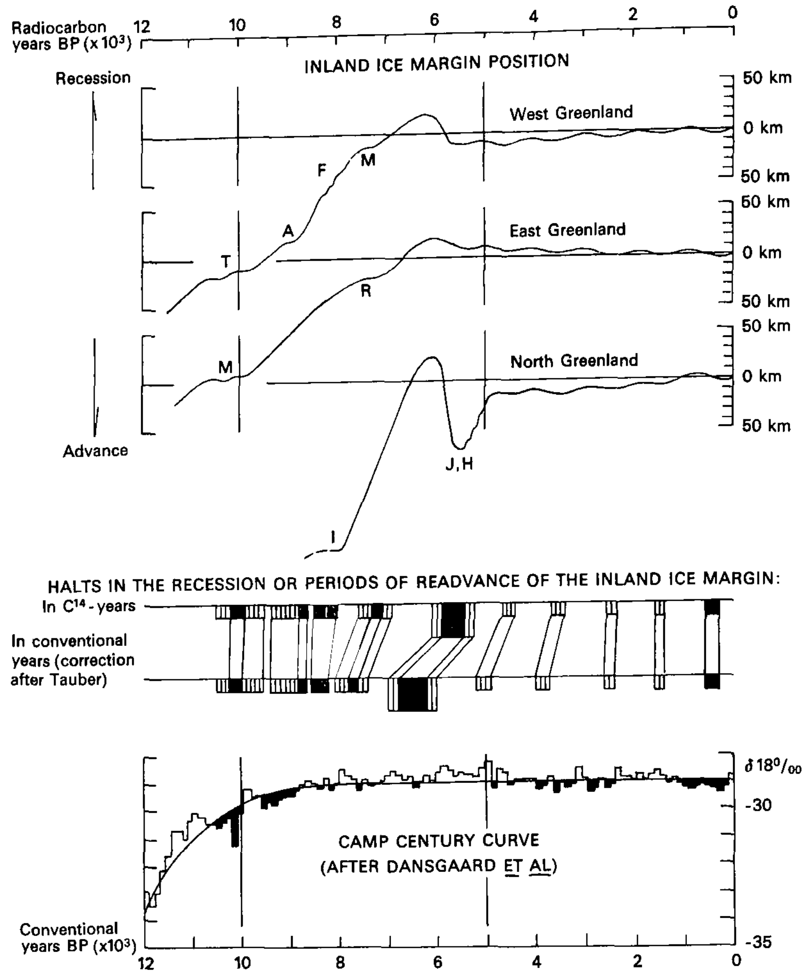


Fig. 31. Top: Generalized curve showing the fluctuations of the Inland Ice margins in West, East and North Greenland.
Letters indicate the following stages in the ice margin position:
West Greenland: T = Taserqat, A = Avatdleq, F = Fjord and M = Mt. Keglen stage.
East Greenland: M = Milne Land and R = Rødefjord stages according to Funder (1972).
North Greenland: I = stages at the entrance of Independence Fjord. J = Jørgen Brønlund Fjord and H = Hall Land stages.

Center: Approximate age of halt periods in the recession, or of readvance of the Inland Ice margin according to the top curves. Black: Dated periods at present; Hatched areas: alternative ages of dated periods or minor readvances of the last 6,000 years.

The uppermost series indicates age of the periods in radiocarbon years, while the lowermost series gives the same periods after the correction proposed by Tauber. The ages thus should be more comparable to those of cold periods of the Camp Century record given below.

Bottom: The Camp Century record according to Dansgaard et al. (1970). Relative cold spells are given in black.

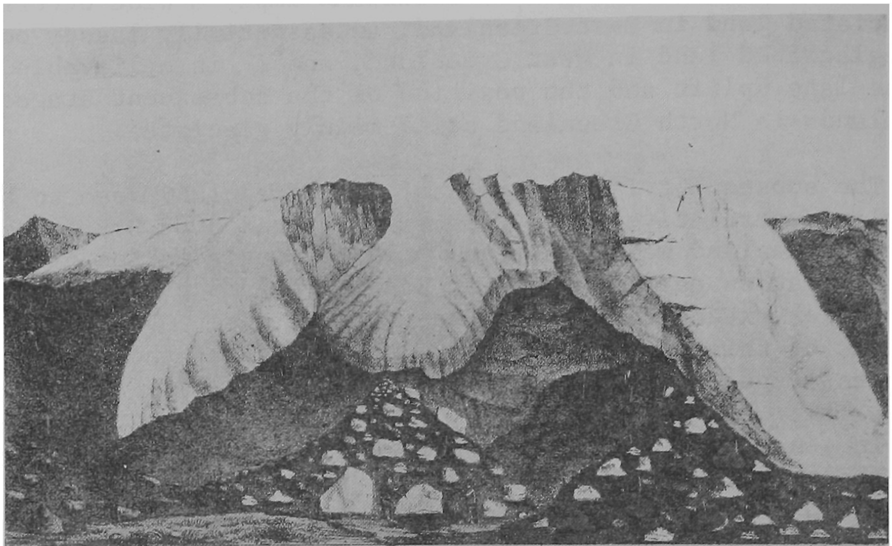
correlative to the Ra-Salpausselkä stage of Scandinavia. At this time, the situation of the stages in Greenland imply a wide occurrence of deglaciated land in East Greenland, an essentially lesser occurrence of deglaciated land in West Greenland, and it is believable in regard to the late uplift and the position of the subsequent stages in North Greenland - a North Greenland still mainly glaciated.

The subsequent fjord stages of West Greenland seem to have North Greenland correlatives. At this time (around 8000 B.P.) essential parts of West Greenland were ice-free while the deglaciation of North Greenland was still at an early stage. In East Greenland the deglaciation may have had an extent close to present conditions. Outside Greenland the moraines of this stage seem very close in age to those of the Cochrane-Cockburn events in Canada.

In general it must be presumed that the recession of the Inland Ice from the coastland in West and East Greenland began 10,000-15,000 years ago and that in North Greenland it began somewhat later. Furthermore, that the recession occurred with a rate of recession of only a few km/100 years and that it was interrupted by short halts or small readvances. The time of general recession continued until the Inland Ice approximately 6000 years ago had an extent less than at present.

Around 6000 years (or possibly between 6000 and 5000 years) ago the period of general recession was locally interrupted by a readvance, especially pronounced in North Greenland but also traceable in West Greenland. It is believed that the readvance reflects a drastic change in the mass balance of the Inland Ice (Weidick, 1971b). Since then the Inland Ice margin as well as the local glaciers have had several minor fluctuations.

The last readvance was presumably initiated in the 16th century and was succeeded by a recession, as in coastal West Greenland and in East Greenland this began in the 18th or 19th century whilst the recession first began after 1900 in the inland parts of West Greenland (Fig. 32) and after 1920 in essential parts of North Greenland (Davies and Krinsley, 1962; Weidick, 1968a). Since about 1940 expansion of a few glacier lobes has been reported (Weidick, 1968a).



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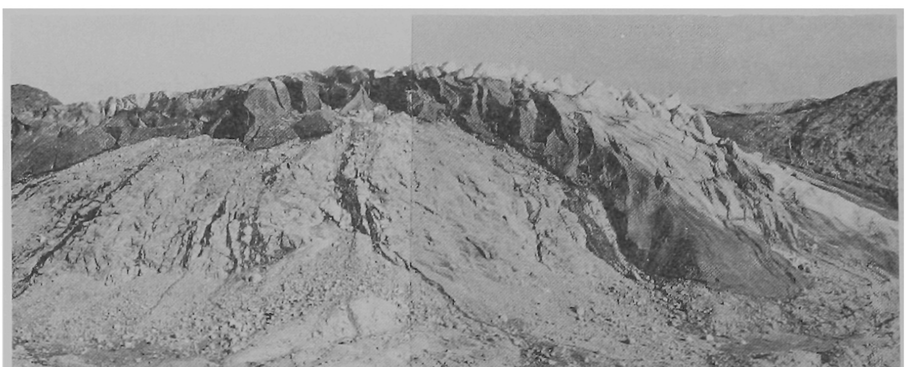

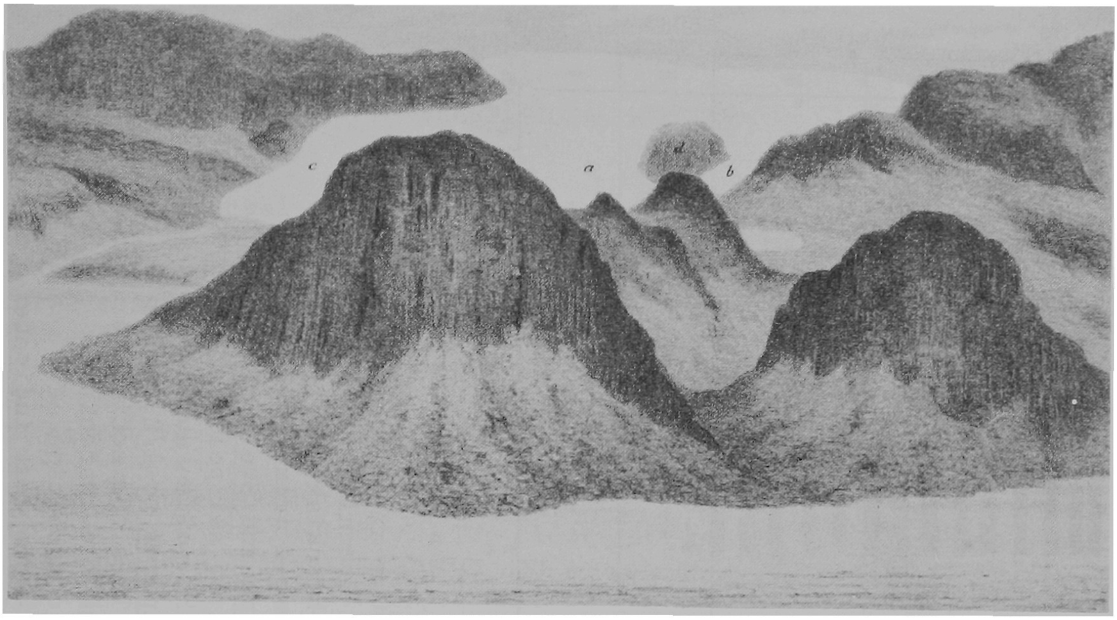

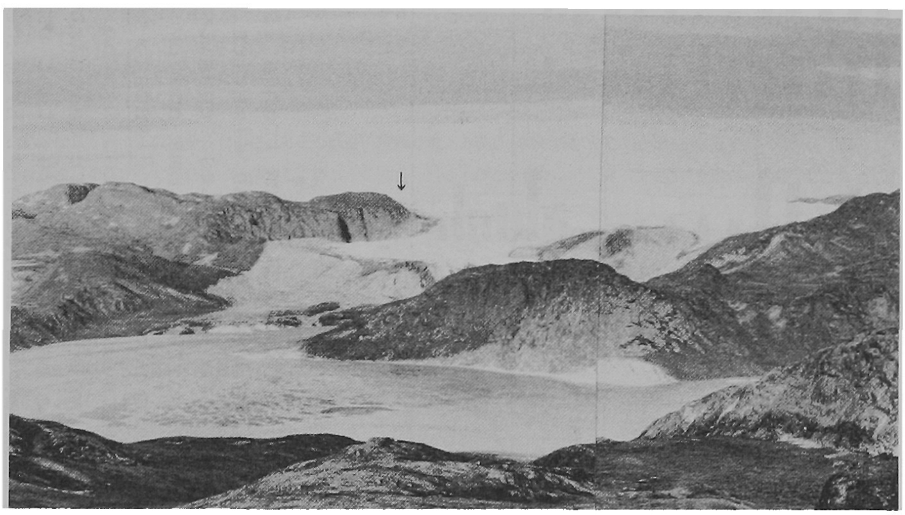

Fig. 32. The Inland Ice lobe at the head of Pâkitsup ilordlia (Qingua kujatdleq).
a) Drawn by Rink around 1850 (Rink, 1857, p. 14).
b) The same glacier seen from Rink's position 23.07.1961 (Photograph, Weidick).
c) Sketch by Hammer drawn in 1883 (Hammer, 1889, table III). The lobe marked "c" is the area glaciated since 1850.
d) The same glacier shown from Hammer's position 21.07.1961 (Photograph, Weidick). Arrow indicates the area where twigs are found in a shear moraine on the Inland Ice margin. The twigs were dated at A.D. 1665 ± 100 (I-5418) which must indicate the time when the glacier was under initial expansion in historical time.

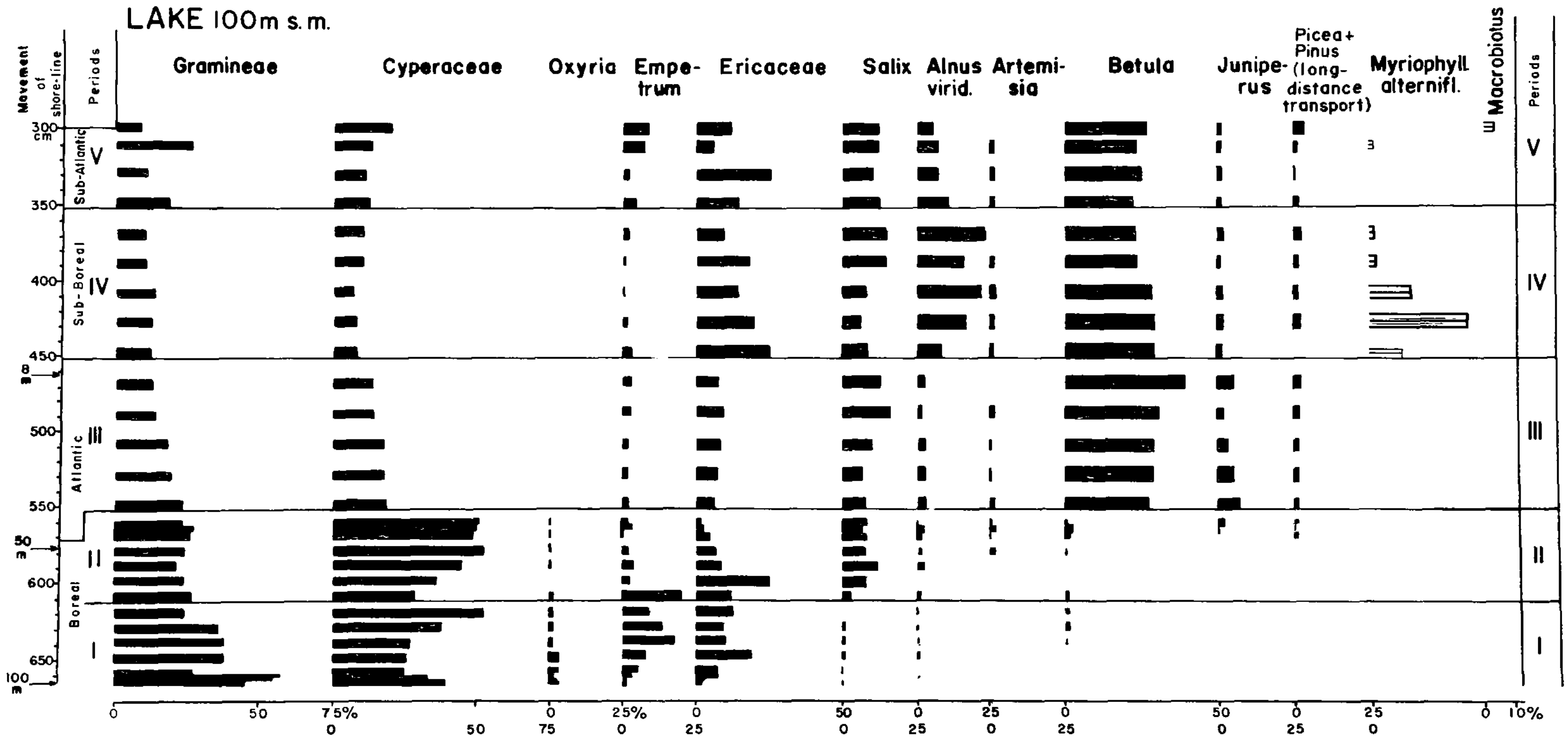

Fig. 33. Composite pollen diagram from a lake 100 m above recent sea level in Kapisigdlit area, Godthåb district, West Greenland. The lowermost analyses are from samples deposited in brackish water, as indicated by their saltwater diatoms and by the ample occurrence of pollen from seashore plants (e.g. Plantago maritima and Stellaria humifosa). The basis of calculation is the total of all pollen, except that of waterplants. Before calculation the pollen number of Betula and Alnus are reduced to 1/4, that of Juniperus to 1/2, while the number is doubled in the case of Ericaceae. This correction is made on the basis of a detailed comparison between the content of recent gyttja in two lakes of the region and the composition of adjoining vegetation areas. The arrows on the extreme left indicate the place in the diagram where 3 different lakes were isolated from the fjord; the heights above recent sea levels of these lakes are 100, 59, and 8 m, respectively (text and illustration according to Iversen, 1953).

VEGETATIONAL HISTORY

Greenland palynological investigations were initiated by Iversen in the early 1930's around Kangersuneq in the innermost parts of Godthåbsfjord in West Greenland (Iversen, 1934, 1953). He obtained cores spanning the longest time yet covered by this method from three lakes at 8, 50 and 100 m a.s.l., respectively. In the cores Iversen discerned stratigraphic zones from early Boreal to Sub-Atlantic times. That these zones correlate reasonably with the Scandinavian zones seems to be confirmed by subsequent C^{14} dating of peat and shells. Besides giving the times of lake-sea isolations, the occurrence of several pollen in the lower part of the sequence indicates temperatures comparable to those of today. The older part of the Boreal period was characterized by lack of scrub and a younger part (period II in Fig. 33) began with a sudden appearance of _Salix_. The Atlantic period (III) was initiated by the migration of _Betula nana_ and _Juniperus_ into the area, but the following Sub-Boreal period is regarded as the climatic optimum in a vegetational aspect because of the immigration of alder. The Sub-Atlantic period was characterized by a recession of alder scrub which was replaced by _Empetrum-Vaccinium_ heath.

Iversen's investigations were for nearly 20 years the only published data covering most of the Holocene. A subsequent treatment covering sites from the Kap Farvel area in South Greenland through West Greenland to Peary Land in North Greenland was given by Fredskild (1973). Fredskild states that on the outer coast of South Greenland, the first vegetation arrived at 10,000 years B.P. and that this was a pioneer vegetation of arctic plants in which no southern species were found. Here in the outer coast of South Greenland, _Salix_ and _Juniperus_ immigrated at 7200 B.P. and _Betula glandulosa_ at 3800 B.P. The Hypsithermal is at 5300-2200 B.P. In the inland region of South Greenland with more continental climate, _Salix_ immigrates at 8000 B.P., _Juniperus_ at 7000, _Betula glandulosa_ at 4500-4000 and _Betula pubescens_ at 3600 B.P. Such a development was also found in the interior parts of Godthåbsfjord. In North Greenland, the vegetation has changed little in the last 5000 years, though with indications of relatively dry periods at 4000-2100 and 1500-900 B.P.

For the youngest Holocene, investigations of the vegetational development is closely connected with archeological work and has been described from several localities by Fredskild. In West Greenland such investigations were made at Iversen's 8 m lake in the Godthåb district and at Qagssiarssuk in South Greenland, and at both localities the investigations were concerned with changes of the flora caused by Norse settlement and subsequent disappearance in the Middle Ages (Fredskild, 1967, 1969b).

Farther north, palynological investigations of the Sermermiut locality near Jakobshavn revealed sequences of peat overlying marine deposits. The investigations covered the last 3500 years and the radiocarbon-dated events showed agreement with the re-growth of European raised bogs about 700-500 B.C. and A.D. 400 (RY III and RY II). While dry conditions prevailed in the beginning of the sequence (overlying the marine silt), more humid conditions developed after 500 B.C. (Fredskild, 1967), and especially between A.D. 500 and A.D. 1000 (Larsen and Meldgaard, 1958).

In North Greenland investigations at Jørgen Brønlund Fjord (Fredskild, 1969a) indicated that after the last advance of the glaciers (see p. 88) the fjord became open and a humid climate prevailed during the paleo-Eskimo settlement (see p. 99). About 2500 years ago the fjords and the polar basin became permanently frozen and the local climate became more arid.

The temperature development revealed by pollen analyses agrees to some extent with the climatic development indicated by other data (see Fig. 31), although the Sub-Boreal climatic optimum of Kapisigdlit at Kangersuneq does not seem to be substantiated by data from other regions. The data do not allow conclusions about humid and dry periods for greater areas (cf. p. 40).

ARCHEOLOGY

The cultural prehistory and the history of Greenland can be divided into three phases which were partly contemporaneous: the Eskimo cultures, the Norse cultures and the colonial period. In the short account given here the greatest emphasis will be upon Eskimo cultures which are the oldest and of greatest stratigraphic significance. Detailed references to archeological works on the Eskimo cultures can be found in Birket-Smith (1961), Knuth (1958, 1967) and Bandi (1969); and references regarding Norse cultures in Roussell (1941), Meldgaard (1965) and Gad (1970). The different cultures will be referred to in chronological order. The ages of the Eskimo cultures are based on dates published by Tauber (1960a, 1961, 1962).

Independence I (4100 - 3700 B.P.) and Sarqaq (3500 - 2700 B.P.) Paleo-Eskimo cultures

Evidence of the Independence I culture is found along the northeast coasts of Greenland (Independence Fjord), and the culture supposedly extended along the east coast as far south as the Scoresby Sund area (Knuth, 1967). Evidence of the Sarqaq culture has been found on the west coast of Greenland around Disko Bugt and in the Sukkertoppen and Godthåb districts. The cultures apparently existed between 2000 and 500 B.C. but according to Knuth there was an essential difference between them as the Independence I culture was based on musk ox hunting and the Sarqaq culture on reindeer hunting. The Independence I culture is characterized by rather large burins without any trace of grinding, and the absence of lamps (Larsen and Meldgaard, 1958). Small lamps occur in deposits of the Sarqaq culture, the size of which indicates that they were only for illumination, not for heating.

It is supposed that people of the Independence I culture migrated to Greenland from Canada along the coasts of the Arctic Ocean, while people of the Sarqaq culture came to Greenland across Smith Sound farther to the south.

Geologically, the Independence I culture in North Greenland is confined to marine levels c. 10 m above the present sea level (Knuth, 1964, 1967). The Sarqaq culture in West Greenland seems to have been contemporaneous with a sea level very close to the present.

At Sermermiut in Disko Bugt, the Sarqaq culture layer is separated from the following Dorset culture by a sterile layer. Therefore, the disappearance of the Sarqaq culture may be explained by a possible climatic deterioration around the beginning of Sub-Atlantic time (Larsen and Meldgaard, 1958).

Independence II (around 2600 B.P.) and Dorset (2000 - 1100 B.P.) Paleo-Eskimo cultures

Evidence of the Dorset culture has been described from the Thule area, Melville, Bugt, Disko Bugt and Angmagssalik. It is known that the Norsemen observed dwelling places in South Greenland, and it is supposed that these ruins were from the Dorset people. Distinctive features of the Dorset culture are the absence of bows and arrows. The culture was based on the hunting of walrus, seal and reindeer. In addition to the artifacts, evidence of the Dorset culture persists in legends of the present population.

The Independence II culture in Peary Land has many implements in common with the Dorset culture, but was apparently adapted to the hunting conditions around the shores of the Arctic sea. Geologically, the Independence II culture in Peary Land was contemporaneous with a sea level c. 4 m above the present one. However, at the time of the Dorset culture, the sea level in West Greenland was close to the present one.

Thule-Inugsuk cultures (from A.D. 1100)

The Thule culture is a Neo-Eskimo culture with more ground artifacts than the older cultures. People of the Thule culture came to Greenland from Canada via the Thule area around A.D. 900. From Thule, some Neo-Eskimos migrated through North and East Greenland to Clavering Ø, and others migrated through West and South-East Greenland to Angmagssalik and Scoresby Sund in East Greenland. Clavering Ø seems to have been occupied from the north and Angmagssalik from the south A.D. 1400-1500. The Thule culture is typified by its extreme specialization in the hunting of sea animals (whale and seal) as reflected by their harpoons and boats (kayaks, umiaks). Especially characteristic of this culture in Greenland are the great lamps used for heating the houses. In both West and East Greenland, the Thule culture has persisted, with various modifications, to the present time. On the west coast, the Inugsuk culture (Mathiassen, 1930) was modified by contact with the Norsemen, while on the east coast the special Angmagssalik culture evolved during the 19th century.

Archeological investigations of the Thule culture have supplied information about minor fluctuations of sea level during the recent past. In general, it seems that the west coast and large parts of the east coast have been sinking. A more detailed report from excavations in South Greenland indicates that sinking there started between A.D. 1600 and 1700 (Mathiassen and Holtved, 1936; Bøgvad, 1940).

Norsemen (A.D. 986 - 1500's)

The Norse migration of the Julianehåb district is historically dated to A.D. 985 or 986, and in the following century they spread to the Frederikshåb and Godthåb districts. The historically dated immigration to Godthåbsfjord was used as a reference level in palynological investigations by Iversen (see also p. 97). A large number of surveys of the extent and character of the Norse settlements have been made (e.g. Roussell, 1941; Vebæk, 1956).

It is known that the Norsemen disappeared from Greenland between A.D. 1500 and 1600. Their disappearance has been explained by a variety of causes including climatic fluctuations, Eskimo attacks from the north, failure of crops, political development in Europe, black death in Europe and Biscayan pirates. Possibly several factors together contributed to the extinction of the Norse culture. In any case, many of the Norse ruins are not being eroded by the sea, and Norse ruins are believed to have been buried under glacier ice in medieval times in Tasermiut and northern Sermilik fjords, Julianehåb district (Weidick, 1959).

PERMAFROST AND PERIGLACIAL STRUCTURAL PHENOMENA

Permanently frozen ground (permafrost) is not a necessary prerequisite for the occurrence of periglacial structural phenomena. In the following section, therefore, the distribution of permafrost is treated on a regional basis and separately from a description of the occurrence of periglacial structural phenomena.

Extent of permafrost in Greenland

The Greenland permafrost area can be divided into three zones:

(1) "Continuous permafrost", where, beneath the active layer, there is a permafrost table which is rarely and only locally absent; e.g., under fjords and rivers (Werenskiöld, 1953) or at warm springs.

(2) "Discontinuous permafrost", where relatively large permafrost islands exist, isolated from each other by areas without permafrost.

(3) "Sporadic permafrost" where permafrost rarely occurs, e.g., as lenses under the shaded sides of valleys. These lenses are often less than 2 m thick.

A connection between climate, niveometric balance, glaciological balance and the existence of permafrost must be assumed to exist, but few details are known about this problem. The southern limit of permafrost approximately coincides with a yearly mean temperature between -3° and -6° in Siberia and -4° to -5° in Canada (Jenness, 1949). In previous delimitations of the circumpolar extent of permafrost these values seem to have been extrapolated for Greenland and the limit of continuous permafrost placed at the annual isotherm of c. -4° (cf. Fig. 34).

Black (1950), Brown (1960, 1969) and Legget et al. (1961) discussed the factors which can govern the existence of permafrost. While the generalization can be made that continuous permafrost is limited by annual temperatures several degrees below the freezing point of pure water, other factors influence this relationship such as the geothermal flux, the electrolyte concentration in groundwater, and the topography.

The following summary is based on data compiled by Hansen (1952), with supplementary data taken largely from Greenland Technical Organization reports. These records are compiled in Fig. 34. Most observations were made at localities near sea level and a correction for altitude is therefore unnecessary.

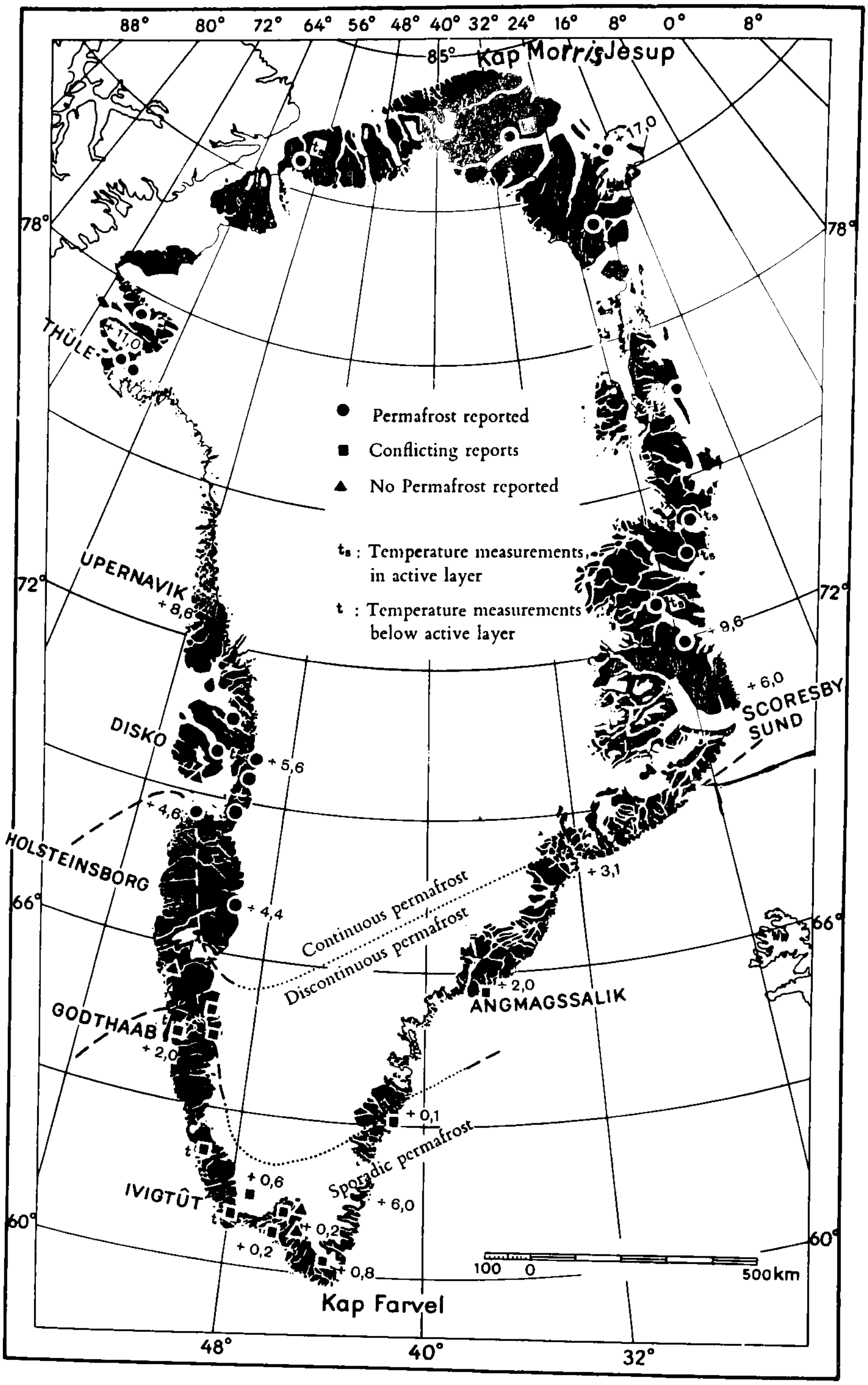

Fig. 34. Extent of permafrost in Greenland. Data based on Hansen (1952) and reports of G.T.O. (Greenland Technical Organisation, Ministry of Greenland). Values shown are mean annual air temperatures. From Weidick (1968a).

The date of observations is important in a discussion of the geographical distribution of permafrost, because climatic fluctuations of the last century have been pronounced in Greenland (Lysgaard, 1949). In West Greenland, annual mean temperature has increased c. 2° in this period, mainly due to increased winter temperature. Apart from heat sources such as rivers, large lakes and the sea, permafrost seems to have the following general distribution:

A zone of permafrost with conflicting reports about its nature covers southern West Greenland as far north as Disko Bugt. Near the Inland Ice margin it seems that permafrost extends farther south than near the coast, but observations are rare. The southern limit of continuous permafrost is placed at a mean annual temperature of c. -5°. South of this limit, permafrost should still be present at higher elevations. With a lapse rate of 0.7°/100 m, the lower limit of continuous permafrost in the Julianehåb and Frederikshåb districts of southern Greenland (mean annual temperature c. 0°) should be c. 600 m a.s.l. Excavations in Grønnedal, Frederikshåb district, showed no permafrost at 350 m a.s.l. (Bøgvad, 1951).

The conflicting reports from southern Greenland are of two types:

(1) Observations made at the same time but not at exactly the same locality, which probably reflect the varying influence of local factors.

(2) Observations made in almost the same excavations but at different times. Reports from archeological excavations in the 1920's and 1930's seem to suggest that permafrost disappeared during this period in the Julianehåb district.

In excavations of the Norse churchyard Herjolfsnes (now Ikigait) Julianehåb district, Nørlund (1924, 1925) found well-preserved medieval garments in frozen earth. From the fact that the garments were interwoven with plant roots, he concluded that the burials occurred at a time when the earth was not frozen. However, Nansen (1925, 1926) pointed out that the shallow depth of the tombs, 30-130 cm, may indicate that permafrost existed here in the Medieval period. Excavations at Brattahlid (now Qagssiarssuk) in 1932 gave more conclusive evidence; no frozen earth was found here in 1932, but was reported in 1896 (Nørlund and Stenberger, 1934).

The thickness of continuous permafrost in West Greenland has been estimated to be c. 150 m at Eqe, Disko Bugt (Boyé, 1950) and c. 500 m at Thule (Roethlisberger, 1961b). In areas with discontinuous or sporadic permafrost, the thickness of the active layer and the depth to the frost-free zone are both between 0.5 and 2.0 m, depending on local exposure and soil type.

It will be seen from Fig. 34 that in East Greenland the limit of continuous permafrost also coincides with the annual isotherm of -5°. The depth of the active layer, or the frost-free zone, varies between 0.5 and 2.0 m, dependent on soil type just as on the west coast, but independent of latitude, so deductions are more limited. However, observations from Skjoldungen and Angmagssalik both seem to indicate a thawing of the earth during the 1930's (Hansen, 1952). The zonation of the permafrost in Greenland is in agreement with observations in Siberia (Ahlmann, 1953) and North America (Black and Barksdale, 1949; Brown, 1960, 1969; Péwé, 1969).

The permafrost is estimated to be c. 360 m thick on Wollaston Forland (74° 30' N, 20° 00' W; Poser, 1932). However, this figure is derived only from the fact that the annual mean temperature is -12° and an assumption that the thermal gradient is 1° /30 m. Poser referred to the success of such calculations in estimating permafrost thickness in Spitsbergen. However, the mean annual temperature at Mesters Vig (72° 52'N, 25° 50'W) is c. -10°, which according to Poser's method of calculation would indicate a permafrost layer c. 300 m thick, a large deviation from the 25-125 m measured thickness (Bondam, 1955; Müller, 1959).

Periglacial structural phenomena

Only the extent of these phenomena will be dealt with here and not their genesis, which has been treated by Müller (1959), Troll (1958) and Washburn (1956a, 1969).

"Non-sorted" solifluction deposits

Structures of this kind are formed by the flow of soil (Fig. 35). Active soil glaciers were observed in 1871 on Disko island, West Greenland (Steenstrup, 1901; Weidick, 1968a). In the basalt areas of West Greenland and East Greenland it is possible to observe numerous soil glaciers on aerial photographs. The soil glaciers seem genetically connected with snow fans and in some degree are related to the type of substratum. In the basalt terrain the moraines have a high content of tufa, which when saturated forms a suspension with a low viscosity. Nevertheless, the soil glacier observed by Steenstrup was capable of transporting large boulders.

Earth streams or boulder streams were reported from North Greenland by Ekblaw (1918) and from East Greenland by Poser (1932) and Stauber (1951).

Fig. 35. Soil glaciers near Ujarasugssuk, island of Disko, West Greenland. Geodetic Institute's route 505 G-V no. 12121 (15.8. 1953). Copyright Geodetic Institute.

Solifluction deposits and structural features are often observed in the uppermost two meters of the ground in excavations for technical installations or foundations for houses. However, descriptions from these excavations are generally not published.

Patterned ground

This category includes stone stripes, stone polygons, soil islands among block fields, and polygonal markings. The principal descriptions of these features are from Arfersiorfik fjord (Jahn, 1946), Disko Bugt (Nieland, 1930; Boyé, 1950), the Thule area (Corte, 1962), Hall Land and Peary Land (Davies, 1961), the area between Wollaston Forland and Scoresby Sund (Poser, 1932; Sørensen, 1935; Bretz, 1935; Washburn, 1967, 1969).

Patterned ground, developed on superficial drift on glacier ice, was described from Thule by Washburn (1956b).

With reference to the vertical distribution of patterned ground Sørensen (1935) reported that active polygonal soil at Ella Ø, East Greenland, could be observed at 0-200 m a.s.l., but above 200 m was less active.

Pingut (sing.: Pingo) and Serssineq (aufeis)

The geological definition of pingut was given by Müller (1959) as: "Conical-shaped hills rising to about 160 ft [50 m], which, in active form, only occur in permafrost regions". Their formation, according to Müller, is due to water of sub- or interpermafrost origin. Müller's descriptions were restricted to pingut in East Greenland but are equally appropriate for those in West Greenland.

In all localities described, pingut are confined to the alluvial plains in large valleys where permafrost is probably thin. From West Greenland, descriptions of pingut in the basalt areas were given by Rosenkrantz (1940, et al., 1942). An analysis of the gas in one of the pingut in Nûgssuaq peninsula showed it to contain 73 percent methane. The gas supposedly originated from Cretaceous bituminous shales under the basalt. From another pingo in the same area Laursen (1950) reported the presence of marine shells, transported upward from Quaternary marine sediments underlying the pingo.

In East Greenland, pingut have been described from the area around Mesters Vig and Traill Ø, where Permo-Carboniferous or Triassic rocks form the substratum (Müller, 1959), from Schuchert Dal (Cruickshank and Colhoun, 1965), from Agassiz Dal (Flint, 1948) and from Hold with Hope, Wollaston Forland and A.P. Olsens Land (Vischer, 1943).

Somewhat related to pingut, a "serssineq" (aufeis Danish: "kildeis") is a great mass of ice formed over a warm spring or any spring which continues to run during the winter. The serssineq can build up to such an extent that it is perennial. A great ice mass of this kind was described by Rink (1857) at Serminguaq (Fig. 36) on the north coast of Nûgssuaq peninsula, West Greenland.

Related to the serssineq and pingut are "hydrolaccoliths" which result from the forcing upwards of a mass of ice, formed over the outlet of a spring, by the pressure of the water beneath. Such formations were described by Ryder (1889) and Porsild (1925) from near Godhavn, West Greenland. Boyé (1950) reported similar features near the Inland Ice margin in Disko Bugt, West Greenland.

Pingo, or pingo-like hills are in the inland parts of West Greenland found as far south as in Nordre Isortoq valley just north of Søndre Strømfjord, i.e. to the present southern border of continuous permafrost. The pingut here may be relics from the medieval climatic deterioration.

Ice wedges are fissures, filled with ice, a few centimeters to 3 m wide and penetrating the uppermost soil cover to a depth of 10 m below the surface (Fig. 37). The origin is explained by contraction of the earth during severe frost with a subsequent water filling of the fissures formed during the contraction in the thaw. Investigations in Alaska indicate that active ice wedges are related to areas with an average annual temperature of -6°. In Greenland they have been observed in Thule (Corte, 1962) and as far south in the West Greenland inland as Søndre Strømfjord air base, where small ice wedges can be seen in the east end of the runway. At least here the southern limit then must be accordant to a mean annual temperature of -4° to -5°. No reports on ice wedges are found from East Greenland.

Weathering

The scattered descriptions of weathering phenomena in Greenland concern physical weathering more than chemical. This may lead to the impression that chemical weathering is of negligible importance as has been given by Hansen (1970) on the basis of his investigations around Søndre Strømfjord and those of Jensen (1965) on clay from the town of Godthåb. However, Tarr (1897a) and later Washburn (1969) from field evidence from North and East Greenland, stress the importance of chemical weathering, although Washburn considers that physical weathering, especially in the form of frost wedging, is the most important agent.

Fig. 36. Serminguaq, north coast of Nûgssuaq peninsula, West Greenland. Arrow shows location of Sermínguaq (Greenlandic: "the little glacier"), which is the greatest area of serssineq reported from Greenland. Geodetic Institute's route 514 K-SV no. 21 (15.7. 1948). Copyright Geodetic Institute.

Fig. 37. Ice wedge (between the hatched white lines) in an excavation near Thule air base, North Greenland. Photograph, August 1956, Weidick.

Chemical weathering

From his detailed studies in the Mesters Vig district Washburn (1969) mentions oxidation, solution and deposition of calcium carbonate, desert varnish, case hardening, exfoliation, cavernous weathering, granular disintegration and development of Arctic brown soil as evidence of chemical weathering, believing especially that the increased absorption by water of carbon dioxide at the low temperature of the thaw season, counteracts to some extent the lowered chemical activity of lower temperatures.

In the southern parts of Greenland granular disintegration seems to occur especially frequently in the areas of nepheline syenite in the Julianehåb district (Jessen, 1896; Ødum, 1927). Typical mushroom and pillar forms occur here, just as in areas where the vegetation cannot root because of fast disintegration of the rock. Overgrazing by sheep may contribute to the vegetation's disappearance in certain areas of southern Greenland as has been suggested by Lund-Andersen et al. (1951) but since the same areas were already described in 1896 as being bare of vegetation, i.e. before the introduction of the sheep farming at the locality, the integral disintegration of the syenite must be the main reason for the formation of the bare areas.

Disintegration of gneiss is also known farther north in West Greenland (Birket-Smith, 1928), being especially described from the Søndre Strømfjord area (Nordenskiöld, 1914; Belknap, 1941; Böcher, 1949), Disko Bugt (Boyé, 1950) and the Umanak district (Rink, 1857; Drygalski, 1897).

In northern West Greenland and North Greenland the influence of chemical weathering has been mentioned by Paterson (1951) and Corte (1962) from areas in the west while from North-East Greenland (cf. p. 53) the description by Davies and Krinsley of the caves in the Cambro-Silurian limestones at Centrum Sø must illustrate the importance of this agent. Farther south in North-East Greenland chemical weathering has been mentioned by Bronner (1948) and Flint (1948) from the areas of the fjord region (i.e. Kejser Franz Josephs Fjord-Kong Oscars Fjord) and from the Mesters Vig area by Washburn (1969).

Related to the chemical weathering is soil genesis, which has been investigated in West Greenland (Bøcher, 1949; Holowaychuk and Everett, 1972) as well as in North Greenland (Tedrow, 1970) and East Greenland (Ugolini, 1966).

Physical weathering

There is now little doubt about the importance of the role of nivation and periglacial processes in the transport of material and in the formation of the details of the present ice-free landscapes of Greenland. This has been discussed by Ekblaw (1918), Poser (1932 and Boyé (1950) and the processes involved were investigated in test sites in Thule (Corte, 1962) and especially in the Mesters Vig area (Washburn, 1967, 1969). The thorough investigations on weathering, frost action and mass wasting in the Mesters Vig area led Washburn to the conclusion that frost wedging is probably the only widespread purely physical weathering process operating. Examples on small-scale nivation are related to solifluction, and its description and numerous examples from Greenland can be found in the sources cited in the foregoing pages.

Of larger scale features most interest must be attached to the widespread block-fields on elevated plateaus (e.g. Drygalski, 1897; Nordenskiöld, 1914; Washburn, 1969) where blocks of local rock can completely mantle the plateaus. Washburn stressed the transition to common talus through block slopes and considers most frost wedging to post-date the last ice cover in the Mesters Vig area. The same is true for West Greenland where, however, there also seem to be transitions from block-fields and block glaciers to common block moraines (Weidick, 1968a). It is possible that parts of the plateaus of the outer coast were ice-free (cf. p. 60) at least during the last glaciation and therefore also that the boulder fields here are of greater age than usual (cf. the nunatak theory, p. 64).

On a regional scale nivation appears to be especially intensive in basalt terrain (Fig. 38), here resulting in the formation of cirques which do not show signs of later glacial erosion, whereas the nivation in the gneiss areas seems to be a slower process, where the development of genuine cirques stretches over more glaciation cycles (cf. pp. 53-66).

Eolian features

Eolian features, especially abrasion features, are strongly connected to the inner, relatively arid parts of the coastland close to the Inland Ice margin, whereas depositional features are more widely spread over the coastal areas.

Special eolian features, attached to the Inland Ice, are mentioned on pp. 19-32.

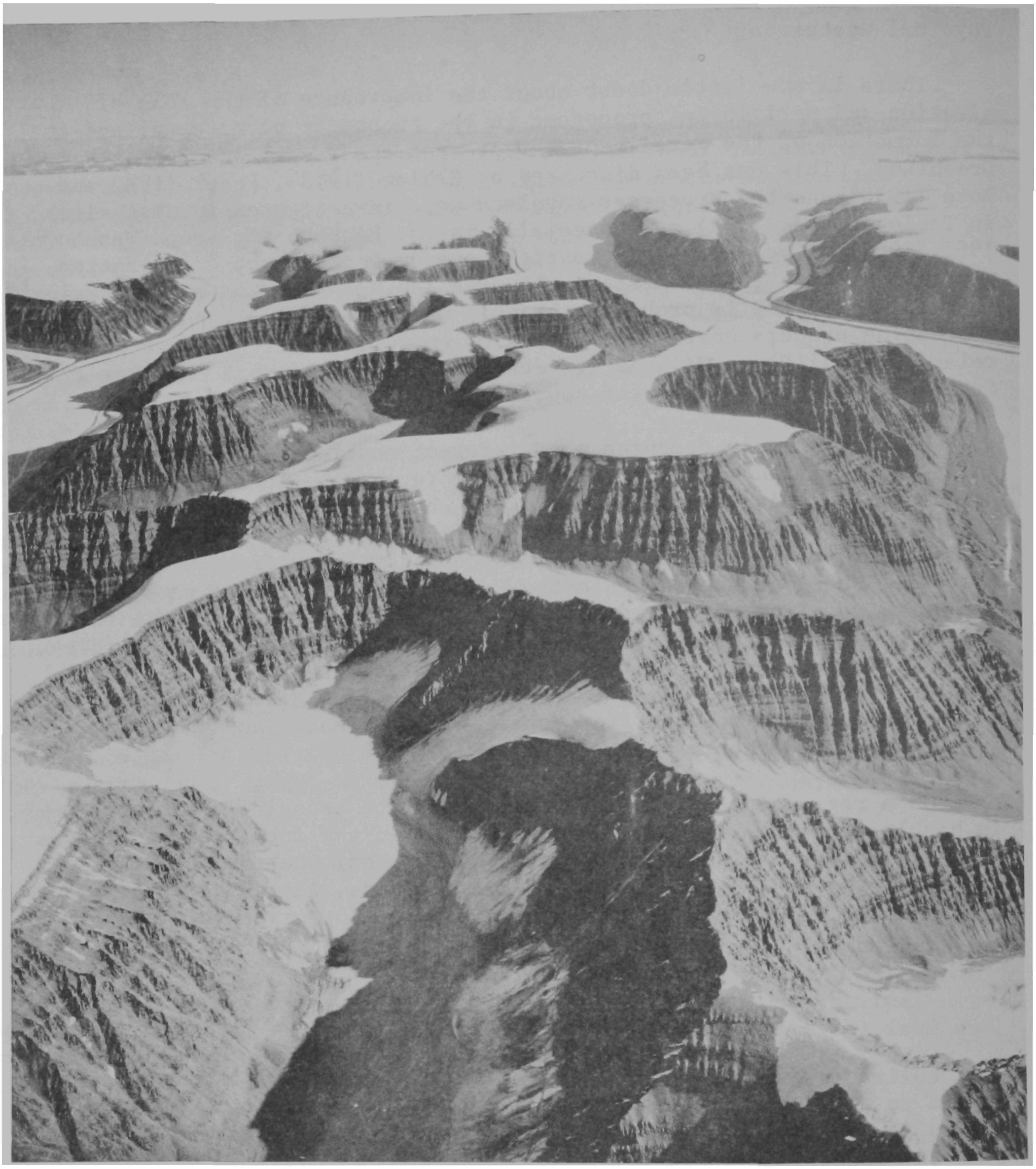

Fig. 38. Cirques in basalt area. Kong Christian IX's Land, East Greenland. Geodetic Institute's route 661 B-NV no. 10650. Copyright Geodetic Institute.

Abrasion features

Wind erosion features occur both in earth and vegetation (stripping and deflation) and in rocks (cavernous weathering, dreikanter).

Wind erosion of soil is known from the Julianehåb district, where the deflation of cultivated earth is a problem for the farmers, here occurring connected with the strong föhn winds (nigeq's) in the inner parts of the district. Farther north stripping of vegetation due to the same kind of winds is reported from Søndre Strømfjord (Jensen, 1889) and Disko Bugt (Boyé, 1950). Allied to the effects of wind is the restriction of lichen growth to the leeside of boulders in the Inland Ice marginal area (Jensen, 1889).

Deflation plains and desert pavement have been described from Julianehåb district (Norlund and Stenberger, 1934), Godthåb district (Roussell, 1941), Søndre Strømfjord (Bøcher, 1949) and Disko Bugt (Boyé, 1950) in West Greenland, and in North-East Greenland from Wollaston Forland (Poser, 1932) and Sandodden near Kap Berghaus (Flint, 1948).

Cavernous weathering occurs especially in sandstone but also in other rock types and is described in West Greenland from Søndre Strømfjord (Belknap, 1941) and Disko Bugt (Boyé, 1950); from the Thule area in North Greenland (Davies et al., 1963); and from Hochstetter Forland (Teichert, 1933) and Scoresby Sund (Teichert, 1935) in East Greenland. Furthermore, the phenomenon is reported in the Mesters Vig area by Washburn (1969), as group cavern formation under chemical weathering, though it is added that deflation is an important accessory process in removing the debris.

In Peary Land an attempt to measure the magnitude of wind erosion has been made by Troelsen (1952b), in an area in Jørgen Brønlund Fjord where ventifacts are frequent. The experiment was conducted over half a year by exposure of blocks of material with hardness 3, and it showed that the wind erosion reached full strength at 1 centimeter above the ground. It was not possible to state whether hard snow contributed to the corrasion. The hardness of snow increases from 2-3 at -15° to 6 at -78.5° (Koch and Wegener, 1930; Teichert, 1939) and it is therefore probable that in parts of northern Greenland it contributes to corrasion.

Depositional features

Deposits of loess have been described from the Søndre Strømfjord area (Nordenskiöld, 1914; Hobbs, 1931; Bøcher, 1959; Hansen, 1970). The cumulative grain-size curves for the Greenland loess given by Hansen are close to those given by Langway and are shown here in Fig. 9.

Farther to the north in West Greenland the present deposition of loess around the Nordre Isortoq river was described by Jensen (1881), where the windblown dust gives the whole area a gray color. The source areas for the eolian deposits are the outwash plains of the great mud-carrying rivers (isortoq Greenlandic "the muddy"). Measurement of the amount of the suspended material in these rivers has been made at the head of Søndre Strømfjord (Jensen, 1881; Bauer et al., 1968b), at Nordre Isortoq (Jensen, 1889) and in the southern branch of Nordre Strømfjord (Jensen, 1889). The values given range between 9,744 g/m^3 (Nordre Isortoq) and 118 g/m^3 (Søndre Strømfjord).

Loess deposits have also been described from Wollaston Forland in North-East Greenland (Poser, 1932).

Dunes are found in the outer coasts, especially where the terrain is low and undulating and where there is morainic material present. In West Greenland their greatest development seems to be around Frederikshåbs Isblink and in Disko Bugt (Porsild, 1902; Krueger, 1928) and in North-East Greenland dune areas are reported from Jameson Land (Hartz, 1896; Hartz and Kruse, 1911).

Inland dunes, near the Inland Ice margin, are more frequent than the marine dunes mentioned above. Their existence is due to the dry climate, the strong catabatic winds from the Inland Ice and the abundant supply of fine-grained material from the outwash plains and moraines. Areas of this kind are described in West Greenland from the Julianehåb district (Jespersen, 1912; Weidick, 1963), from the interior parts of Søndre Strømfjord (Hobbs, 1931; Bøcher, 1949), in North Greenland from Peary Land (Fristrup, 1953) and in North-East Greenland from Kejser Franz Josephs Fjord and Scoresby Sund (Storgaard, 1928; Flint, 1948; Oosting, 1948). In addition to those already described the mapping of the Quaternary deposits has revealed many areas of inland dunes (Weidick, 1971a).

Salt lakes and formation of evaporites

The distribution of salt lakes is restricted to the regions near the margin of the Inland Ice with a severely continental climate. The salt lakes at the head of Søndre Strømfjord, West Greenland, have been the subject of several investigations (Jensen, 1889; Bøcher, 1949; Hansen, 1970). Similar features are reported from Peary Land in North Greenland (Fristrup, 1953).

In relatively dry areas of Greenland efflorescence on soil, sediments and bedrock is common. For example, crusts of salts can be observed on many marine deposits in the inner parts of Holsteinsborg and Egedesminde districts and similar crusts have been observed near Mesters Vig, North-East Greenland, where calcium carbonate is common but where also trona, otherwise only reported from warm-arid environments, was found (Washburn, 1969).

WARM SPRINGS AND SULFUR MOUNDS

While hot springs do not seem to have been found anywhere in Greenland, there are several warm springs having temperatures between 62°C and 5°C. Furthermore, numerous "frost-free" springs are found along the coast. Both warm springs and frost-free springs in West Greenland often have the name Ûnartoq (plural: Ûnartut).

The best known example in Greenland is Ûnartoq in Agdluitsoq fjord, Julianehåb district. The spring was first described by Ivar Baardsøn in the Middle of the 14th century and has been described subsequently by visitors to the spring since 1806 (Persoz et al., 1972). Descriptions of the temperature of the spring in these years show that it has been nearly constant since the Middle Ages. The measurements during the last 100 years all give temperatures between 36°C and 41°C. The bubbles coming up through the water consist of nearly pure nitrogen (Jessen, 1896; Persoz et al., 1972).

Frost-free springs have not been described in detail and transitions between them and warm springs probably exist. A list of "warm springs" around Disko Bugt, West Greenland, was given by Rink (1857) and from this region, near Egedesminde, Birket-Smith (1928) described a warm spring with a temperature between 5°C and 6°C.

In East Greenland, warm springs were described from an area in the southernmost parts of Liverpool Land (Pedersen, 1926). Their temperatures range between 6°C and 62°C. Their gas was also found to consist mainly of nitrogen. It is therefore assumed that here and at Ûnartoq, the gas is atmospheric air deprived of its oxygen.

Mounds of gravel and sand, about 10-20 m in diameter with 1-2 m high walls have been found in Jorgen Brønlunds Fjord, Peary Land, North Greenland. The walls and the bottom of the craters were covered with a crust containing gypsum, sulfur, copiapite, pyrite and fibroferrite. The formation of the mounds was ascribed by Pauly to the oxidation of pyritic ores under the mounds (Troelsen, 1949a, b).

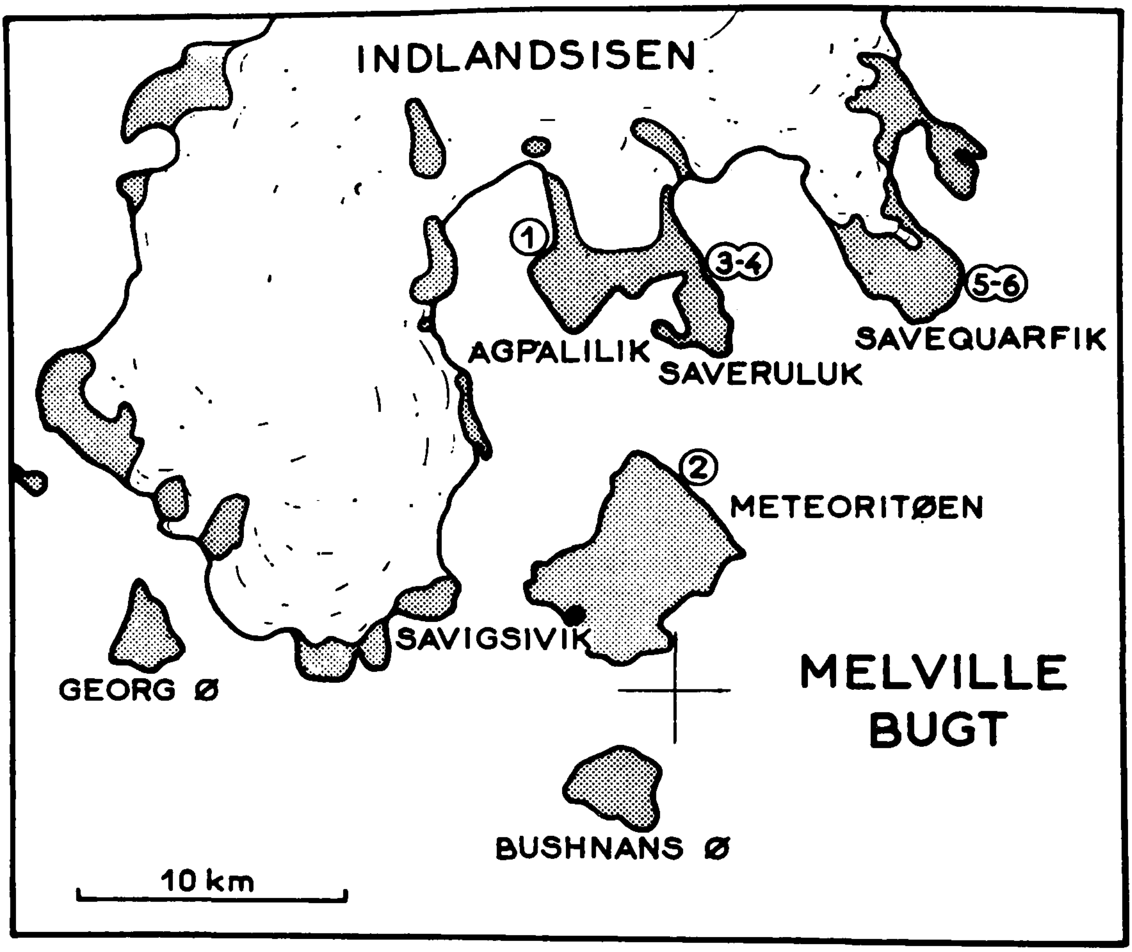

Fig. 39. The original locations of the meteorites around Savigsivik outpost (at approximately 70° 02' N. Lat., 62° 52' W. Long.): (1) "Agpalilik", (2) "Ahnigito", (3) "Woman", (4) "Dog", (5) "Savik I" and (6) "Savik II". From Buchwald (1963). The cross indicates 76° N Lat. and 65° western Long.

Several iron meteorites, all found in the Thule-Kap York area, western North Greenland (Fig. 39), seem to have originated from a meteoric shower during the Quaternary. The date is based on the short distance of the localities from the Inland Ice margin, their superficial situation on the block fields near the present beach, and the freshness of their "thumb marked" surfaces, formed during passage through the atmosphere.

The meteorites have the characteristics shown in Table 9 (from Bøggild, 1930; Buchwald, 1961, 1963).

Table 9. Iron meteorites from Greenland

Name	Weight, kg	Fe%	Ni%
"Ahnigito"	31,000	91.48	7.79
"Woman"	3,000	91.47	7.78
"Dog"	400	90.99	8.27
"Savik"	3,400	not. det.	7.27
"Thule"	48.6	90.7	8.13

Bøggild supposed that the boulders discovered first (i.e. "Ahnigito" to "Savik") were pieces from one single meteorite shower. The report by Peary of a legend among the Eskimos about a meteorite fall can scarcely be taken as a proof of the young age of the meteorites because it cannot be excluded that the Eskimos heard about the possible meteoric origin of the iron from early explorers.

Most of the boulders were long known by the Eskimos who used the iron for their implements. The first scientist to see the boulders "Ahnigito", "Woman" and "Dog" was Peary who transported them to New York in the years after 1894.

On the island Savigsivik, only a few kilometers from the other localities, a new boulder, "Savik" was found in 1913 by Qitlugtoq who led Knud Rasmussen to it (Buchwald, 1964). This boulder was transported to Copenhagen in 1925. Another meteorite, "Thule", was found in 1955 on the south side of Harald Moltke Bræ on a nunatak near the margin of the Inland Ice. The discovery was made by J. Colonel and the boulder described by Buchwald (1961).

Two meteorites have subsequently been found close to the locality of the original four. One of these, "Savik II", weighing c. 8 kg was found by Suerssaq in 1961, the other, "Agpalilik", weighing c. 15,000 kg, was found by Buchwald in 1963. In addition a small meteorite, weighing 0.25 kg, was discovered on Northumberland Ø c. 250 km NNW of the Savigsivik area. This meteorite may possibly belong to the same swarm, but could have been transported by Eskimos.

Possible traces of a meteoric shower south of Peary Land, North Greenland, were reported by Ellitsgaard-Rasmussen (1954).

A field closely connected with research on meteorites is the study of "spherules" or "granules" in the Inland Ice. Nordenskiöld (1874) considered that meteoric dust was present in cryoconite holes in the marginal areas of the Inland Ice, which later was demonstrated by de Quervain and Mercanton (1925). In the marginal areas of the Inland Ice, as on most of the world's surface, it is difficult to separate the sediments of terrestrial origin from those of cosmic origin. However, the "spherules" of the Inland Ice cores (cf. p.42) are possibly of cosmic origin.

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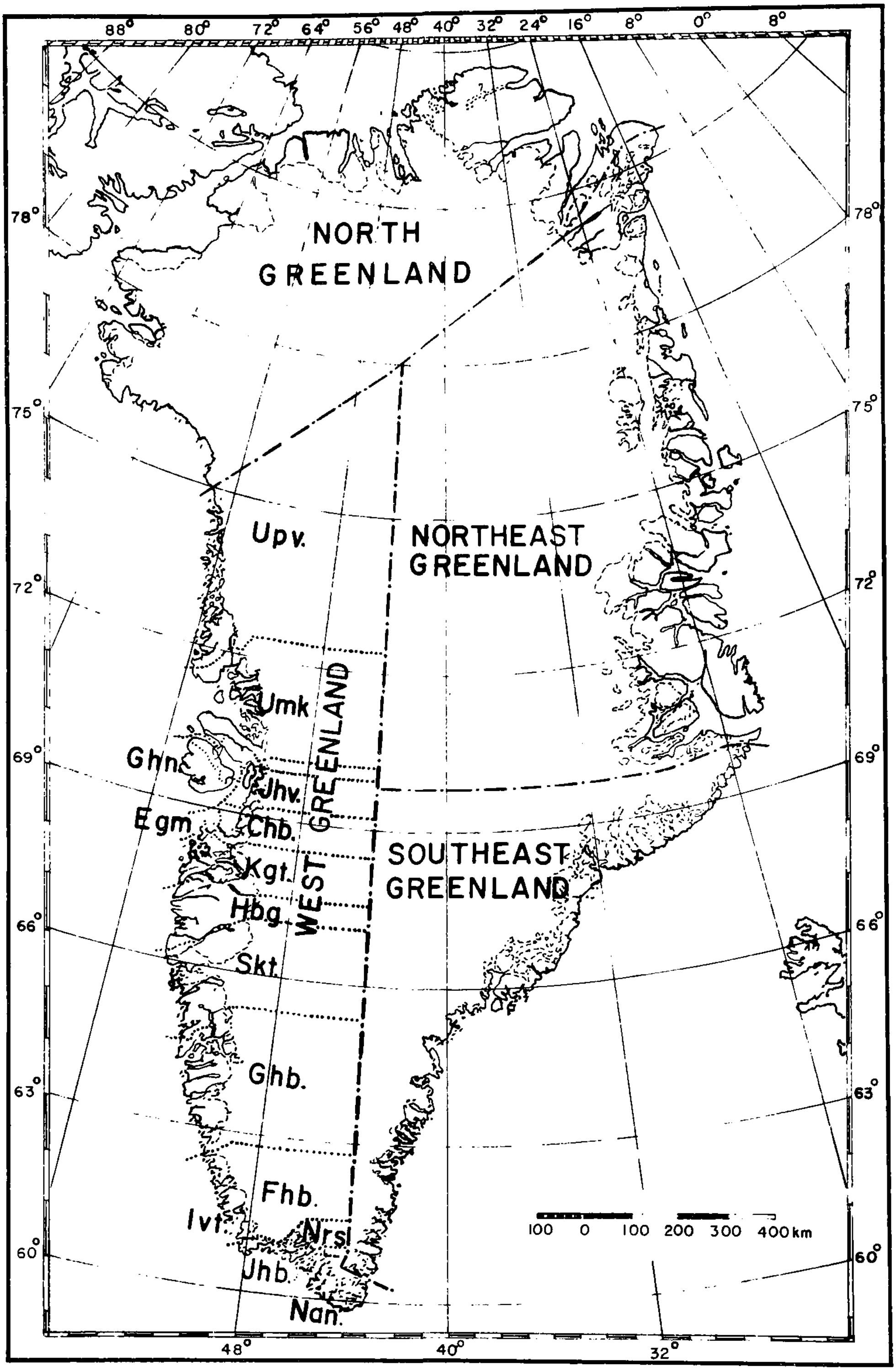

Fig. 40. Regional division of Greenland used in the text (After General Index of "Meddelelser om Grønland").

ABBREVIATIONS

Provinces: N Grl = North Greenland
NE Grl = North-East Greenland
SE Grl = South-East Greenland
W Grl = West Greenland

Local areas of West Greenland:

Chb = Christianshåb
Egm = Egedesminde
Fhb = Frederikshåb
Ghb = Godthåb
Ghn = Godhavn
Hbg = Holsteinsborg
Ivt = Ivigtut
Jhb = Julianehåb
Jhv = Jakobshvn
Kgt = Kangâmiut
Nan = Nanortalik
Nrs = Narssaq
Skt = Sukkertoppen
Umk = Umanak
Upv = Upernavik
Vaj = Vajgat

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